Geomorphology and Hydrogeology of the Edwards Plateau Karst, Central Texas

Ernst H. Kastning

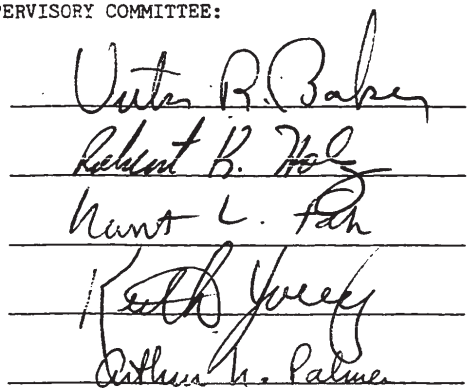
Texas Speleological Survey
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GEOMORPHOLOGY AND HYDROGEOLOGY

OF THE EDWARDS PLATEAU KARST,

CENTRAL TEXAS

APPROVED BY SUPERVISORY COMMITTEE:



This is a reformatted version of the compete PhD dissertation done under the supervision of Victor R. Baker at the University of Texas at Austin in 1983.

Texas Speleological Survey
Pickle Research Campus 176-R4000
The University of Texas at Austin
Austin, Texas 78712
www.texasspeleologicalsurvey.org

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This dissertation is dedicated to
KAREN and KASS
who have given immeasurably of
themselves during the years of
study, research, thinking, writing,
drafting, and earning a living.

CONTENTS

Foreward to the TSS edition 6
Abstract 7
Detailed Contents of Text 9
List of Figures 15
List of Tables 19
List of Plates 19
Preface 21

Part I—General Introduction and Physical Setting
1. General Introduction 25
2. Geological Setting of the Edwards Plateau 37
3. Hydrogeologic Setting of the Edwards Plateau 43

Part II—Investigations of Selected Areas
4. Distribution of Karst and Caves 55
5. Langtry Area 57
6. Sonora Area 71
7. Menard Area 87
8. Bend Area 93
9. Burnet Area 137
10. New Braunfels Area 149
11. Austin–San Marcus Area 165
12. Georgetown Area 179
13. Granitic Pseudokarst in the Llano Basin 187

Part III—Synthesis, Conclusions, and Recommendations
14. Regional Synthesis and Conclusions 201
15. Recommendations for Further Study 205

Appendices
A. Definitions of Karst and Pseudokarst 213
B. Geologic Tables 214
C. Caves of the Edwards Plateau 233
D. Techniques of Mapping Caves 238
E. Evaluation of Remotely Sensed Imagery in Mapping Fractures . 241
F. Drainage Density of Carbonate Terrane 244

References Cited 248
Plates 275

FOREWORD TO THE TEXAS SPELEOLOGICAL SURVEY REISSUE April 2015

This publication represents all the information originally included in Ernst Kastning's 1983 dissertation, published by University Microfilms International (now ProQuest), and still available there. Much of the data in this text is very dated and should not be considered the state of the art in cave research, but this dissertation was a landmark in the history of Texas caving and richly deserves to be accessible, and so the TSS decided to offer a new version of the text and figures, without replicating the original format.

That decision led to an adventure in digital media, because the reproduction quality of the many figures and photographs included in the original reproductions of that dissertation was poor at best. The digital original still existed on a 12-inch magnetic tape reel, but there was no equipment available that could read it. That problem was solved when Katie Arens and Bill Russell found a data recovery service just outside Washington, DC, that "used algorithms" to bail out over 90% of the original text, including some fiendishly complex tables. Katie Arens had to retype the remainder (lost to digital degradation), remove mystery characters, and reformat the text. Jim Kennedy picked up the ball and worked with Ernst Kastning to find and rescan most of the original figures (a very few have vanished, and so the versions from the published dissertations were used); however, the original photographs included had disappeared. Katie Arens then stepped in again to borrow one of the original copies of the dissertation from the University of Texas library, in order to rescan the original photos and clean up some of the original figure scans. The original color photographs were badly faded; their appearance has been improved, but colors are not accurate. Finally, Bill Mixon stepped in to format the text for PDF republication, with his inimitable eye for readability.

Thanks to all for making this possible. Enjoy.

ABSTRACT

The Edwards Plateau is one of the largest contiguous karstic regions of the United States; yet, its geomorphic evolution has previously received little systematic study. The objectives of this investigation are to (1) describe the physical characteristics of karstic features, (2) determine which geomorphic and hydrogeologic controls and processes have governed their development, and (3) relate genesis of karst spatially and chronologically to the geomorphic evolution of the Edwards Plateau.

Caves provide a record of the evolution of karst and the development of major carbonate aquifers. Some controls on speleogenesis are pervasive; nevertheless, many caves exhibit characteristics that suggest a strong influence of local factors

Lithostratigraphic factors have influenced the development of chambers and passages. Solutional conduits were guided by variations in calcitic and dolomitic content, thickness of strata, and frequency of bedding-plane partings. This is particularly true of the Glen Rose Formation, limestone beds of the Edwards Group, and the Gorman Formation, the three principal cave-forming units. Caves were developed preferentially along fractures associated with regional structural elements, including the Devils River Uplift, Llano Uplift, San Marcos Arch, and Balcones Fault Zone. Locally, faults have modified passages.

Evolution of flowpaths within major aquifers has governed the morphology, distribution, and orientation of caves. Patterns of caves vary markedly and include isolated cavities, interconnected cavities, single conduits, branching conduits, and network, anastomotic, and spongework varieties of mazes. These suggest a diversity of hydrogeologic controls and, in many places, a strong influence of the evolving topography. Most caves of the plateau are relict features, abandoned by groundwater as dissection of the area progressed. They are situated well above floors of valleys and represent early stages of speleogenesis. However, where streams have incised deeply along the Balcones Escarpment and at some places in the interior of the plateau, caves are well integrated, at grade with surficial streams at baselevel, and characteristic of later stages of development

Climatic changes in central Texas during the Quaternary are reflected in siliclastic sediments and speleothems in caves. Ample evidence suggests that caves were severely flooded during periods of colder and wetter climates.

DETAILED CONTENTS OF TEXT

PART I
GENERAL INTRODUCTION AND PHYSICAL SETTING

1. GENERAL INTRODUCTION 25
Purpose 25
Statement of Problem 25
Needs for This Study 25
Scope of Present Study 25
Summary of Concepts 25
The Nature of Karst 25
Definition of Karst 25
Terminology of Karst 25
Geomorphology and Hydrogeology of Karstic Terranes 25
Speleogenesis 25
Major Emphases of Present Study 26
Lithostratigraphy 26
Structure 26
Hydrogeology 26
Topographic Evolution 26
Climate 26
Geographic Setting 26
Area of Study 26
Principal Geographic Regions of Central Texas 27
Physiographic Divisions 27
Topography 27
Drainage 28
Resources of Groundwater 28
Soils 29
Previous Investigations and Review of the Literature 29
Existing Data-Base 30
Investigations in the Edwards Plateau 30
Geologic Problems 30
Geomorphic Problems 32
Hydrogeologic and Geochemical Problems 33
Environmental and Engineering Problems 33
Speleological Problems 35
Surficial Karst 35
Paleokarst 35
Nitrate Deposits in Caves 35
Paleontological Deposits. 35
Summary 35

2. GEOLOGIC SETTING OF THE EDWARDS PLATEAU 37
Introduction 37
Stratigraphy 37
Structural Features 37
Summary of the Geologic History of the Edwards Plateau Region 39
Precambrian Era 39
Paleozoic Era 39
Mesozoic Era 42
Cenozoic Era 42

3. HYDROGEOLOGIC SETTING OF THE EDWARDS PLATEAU 43
Introduction 43
Hydrologic Cycle . 43
Groundwater 43
Major Aquifers 43
Edwards-Trinity (Plateau) Aquifer 43
Edwards (Balcones Fault Zone) Aquifer 43
General Hydrology and Geology 43
Structure of the Edwards Aquifer 48
Subdivisions of the Aquifer 48
Porosity 48
Bad-Water Line 49
Relationships between Precipitation and Recharge 49
Relationships between Recharge and Discharge 49
Geochemical Studies of the Edwards Aquifer 51
Evolution of the Edwards Aquifer 51
Trinity Group Aquifer 51
Minor Aquifers . 51
Ellenburger-San Saba Aquifer 51
Hickory Aquifer 52
Other Aquifers 52
Additional Information . 52

PART II
INVESTIGATIONS IN SELECTED AREAS

4. DISTRIBUTION OF KARST AND CAVES 55
Karstic Regions of Texas 55
Distribution of Surficial Karst on the Edwards Plateau 55
Distribution of Caves 55
Selected Areas for Detailed Investigation 55

5. LANGTRY AREA 57
Description of Area 57
Distribution of Caves and Previous Investigations 59
Development of Caves 60
Descriptions of Caves 60
Langtry Lead Cave 60
Emerald Sink 61
Langtry Quarry Cave 61
Langtry Gypsum and Langtry East Gypsum Caves 61
Fisher's Fissure 63
Speleogenesis 64
Lithologic Control. 64
Structural Control 64
Relationship to Fractures 64
Modification by Collapse 67
Hydrogeologic and Topographic Controls 70
Bathyphreatic Development of Caves 70
Chronology of Speleogenesis 70
Summary 70

6. SONORA AREA 71
Description of Area 71
Distribution of Caves and Previous Investigations 71

Development of Caves 71
Descriptions of Caves 71
Caverns of Sonora 71
Felton Cave 75
Silky Cave 75
Speleogenesis 78
Lithologic Control 78
Structural Control 81
Relationship to Fractures 81
Modification by Collapse 83
Hydrogeologic and Topographic Controls 83
Sediments and Speleothems 84
Chronology of Speleogenesis 85
Summary 85

7. MENARD AREA 87
Description of Area 87
Distribution of Caves and Previous Investigations 87
Development of Caves 87
Descriptions of Caves 87
Powell Cave 87
Neel Cave 89
Silver Mine Cave 89
Meteor Hole 89
Lakeside Spring 90
Topographic Relationships 90
Speleogenesis 90
Lithologic Control 90
Structural Control 90
Hydrogeologic Control 91
Development of Mazes 91
Chronology of Speleogenesis 92
Summary 92

8. BEND AREA 93
Description of Area 93
Distribution of Caves and Previous Investigations 93
Development of Caves 94
Descriptions of Caves 97
Gorman Cave 97
Lemons Ranch Cave 97
El Sotanito Grande, Yates Goat Cave, Barnes Bat Cave 100
Clark's Branch Well Cave 101
Gorman Fissure System 102
Bluff-Caves 102
Northwest Caves 108
Gorman Falls Cave 109
Caves in Travertine 109
Caves of the McLarrin Fissure Karst 109
Speleogenesis 117
Lithologic Control 119
Structural Control 127
Relationship to Attitude of Beds 127
Relationship to Fractures 127
Synergetic Effects of Lithology, Structure, and Hydrogeology 127

Topographic Control 129
Origin of Bluff-Caves 131
Origin of Caves with Single Rooms 133
Origin of Pit-Caves 133
Origin of Fissure-like Caves 133
Origin of Caves in Travertine 133
Chronology of Speleogenesis 133
Summary 135

9. BURNET AREA 137
Description of Area 137
Distribution of Caves and Previous Investigations 137
Development of Caves 140
Description of Longhorn Cavern 140
Speleogenesis 144
Lithologic Control 144
Structural Control 144
Relationship to Attitude of Beds 144
Relationship to Fractures 144
Modification by Collapse 145
Hydrogeologic and Topographic Controls 146
Phreatic Phase in Development of the Cave 146
Vadose Phase in Development of the Cave 146
Modification by Floodwater 148
Chronology of Speleogenesis 148
Summary 148

10. NEW BRAUNFELS AREA 149
Description of Area 149
Distribution of Caves and Previous Investigations 149
Development of Caves 149
Descriptions of Caves 149
Natural Bridge Caverns System 149
Bracken Bat Cave 151
Double Decker Cave 153
Speleogenesis 154
Lithologic Control 154
Structural Control 155
Bedding-Plane Partings 158
Joints 158
Control by Faults 158
Correlation of Caves with Surficial Fractures 159
Modification by Collapse 160
Speleothems 163
Chronology of Speleogenesis 163
Summary 164

11. AUSTIN-SAN MARCOS AREA 165
Description of Area 165
Distribution of Caves and Previous Investigations 165
Development of Caves 165
Descriptions of Caves 165

Airman's Cave 165
Cave X 167
Bee Creek Cave 167
Dead Dog Cave No. 2 171
Swimmin' Hole 171
Wonder Cave 175
Ezell's Cave 175
Wimberley Bat Cave 175
Jacobs Well 175
Speleogenesis 176
Lithologic Control 176
Structural Control 176
Hydrogeologic Control 176
Chronology of Speleogenesis 177
Summary 177

12. GEORGETOWN AREA 179
Description of Area 179
Distribution of Caves and Previous Investigations 179
Development of Caves 181
Description of Inner Space Cavern 183
Speleogenesis 183
Lithologic Control 183
Structural Control 184
Relationship to Attitude of Beds 184
Relationship to Fractures 184
Modification by Collapse 185
Hydrogeologic and Topographic Controls 185
Chronology of Speleogenesis 186
Summary 186

13. GRANITIC PSEUDOKARST IN THE LLANO BASIN 187
Introduction 187
Geographic Setting 187
Previous Geomorphic Investigations 187
Geologic Setting 187
Weathering and Erosion of Granite 187
Structural Landforms 189
Granitic Pseudokarst 191
Gnammas 191
Rock Doughnuts 191
Granitrillen and Channels 193
Tafoni, Arches, and Pedestals 193
A-tents 193
Caves Formed by Exfoliation and in Talus 193
Enchanted Rock Cave 196
Fracturing 196
Grusification 197
Mechanical Suffosion 197
Roofing by Accumulation of Talus 197
Summary 197

PART III
SYNTHESIS, CONCLUSIONS, AND RECOMMENDATIONS

14. REGIONAL SYNTHESIS AND CONCLUSIONS 201
Introduction 201
Controls on Development of Caves 201
Lithostratigraphic Factors 201
Structural Factors 201
Attitude of Beds 201
Fractures 201
Localization of Development of Caves 202
Structural Stability and Collapse 202
Hydrogeologic Factors 202
Topographic Factors 202
Climatic Factors 202
Chronology of Development of Caves 203

15. RECOMMENDATIONS FOR FURTHER STUDY 205
Regional Synthesis of Locally Disparate Conditions 205
Suggestions for Future Research 205
Additional Localities 205
Regionwide Geologic Studies 205
Lithostratigraphic Studies 205
Structural Studies 208
Hydrodynamic and Geochemical Studies 208
Quantitative Geomorphic Studies 209
Regional Synthesis and Characterization 209
Contribution to Knowledge of Speleogenesis 209
Applications toward Resources and Environmental Problems 209
Summary 209

APPENDICES

A. DEFINITIONS OF KARST AND PSEUDOKARST 213
Definitions of Karst 213
Definitions of Pseudokarst 213

B. GEOLOGIC TABLES 214
Stratigraphic Units of the Edwards Plateau 214
Geologic Column of Pre-Cretaceous Rocks (Plate 4) 214
Correlation Chart of Cretaceous Rocks (Plate 5) 214
Geologic Map of the Edwards Plateau (Plate 3) 214
Structural Map of the Edwards Plateau (Figure 10) 214

C. CAVES OF THE EDWARDS PLATEAU 233
Cataloging of Caves of the Edwards Plateau 233
Significant Caves of the Edwards Plateau 235
Long and Deep Caves of Texas 236

D. TECHNIQUES OF MAPPING CAVES 238
General Procedures 238
Surveying 238
Cartography 238
Mapping of Caves in the Edwards Plateau 238

E. EVALUATION OF REMOTELY SENSED IMAGERY IN MAPPING FRACTURES IN A CARBONATE TERRANE 241
Introduction 241
Lineaments and Fractures 241
Studies of Lineaments in Texas 241
Analysis of Imagery 241
LANDSAT (ERTS) Band 5 Imagery 530
Color-Infrared Imagery 241
Black-and-White Photomosaic 243
Black-and-White Stereoscopic Imagery 243
Summary 243

F. DRAINAGE DENSITY OF CARBONATE TERRANE 244
Introduction 244
Area of Study 244
Methods of Analyzing Drainage Density 244
Application to Southeastern Edwards Plateau 246

REFERENCES CITED 248

LIST OF FIGURES

1 The area of study encompassing the karst of the Edwards Plateau 28
2 Physiographic subdivisions of the Edwards Plateau 28
3 Drainage basins of major rivers on the Edwards Plateau 28
4 Major aquifers of central Texas 29
5 Minor aquifers of central Texas 29
6 Major associations of soils of the Edwards Plateau. 29
7 Interrelationships among geologic, biologic, and geographic environments 35
8 Generalized geologic section of Cambrian and Lower Ordovician rocks of the Llano Basin. 38
9 Paleogeographic and structural elements of central Texas during the Early Cretaceous and composite map of lithofacies 40
10 Composite map of major structural and tectonic elements of central Texas 41
11 Schematic cross-section of the Edwards Plateau showing components of the hydrologic budget and elements of the aquifers 44
12 Hydrologic setting of the Edwards Aquifer 46
13 Approximate altitude of water levels in the Edwards and Edwards-Trinity Aquifers 47
14 Southwesternmost spring at Comal Springs, New Braunfels, Comal County 48
15 Relationship between annual recharge and precipitation for seven sub-basins in the area of San Antonio 49
16 Hydrographs of Comal, Hueco, and San Marcos Springs 50
17 Karstic regions of Texas 55
18 Geologic map of the Langtry area, Val Verde County 58
19 Geologic section along the regional dip in the vicinity of Langtry. 59
20 Map and profile of Langtry Lead Cave. 60
21 Map and profile of Emerald Sink 61
22 Map and profile of Langtry Quarry Cave 62
23 Maps of Langtry Gypsum Cave and Langtry East Gypsum Cave 63
24 Map and profile of Fisher's Fissure 63
25 Cross-sectional sketch of the entrance to Langtry Quarry Cave 64
26 Rose-diagrams of linear segments of passages and mapped photolineaments, Langtry area . . . 65
27 Map of lineaments and large-scale features of solution and subsidence in the Langtry area 66
28 Compressional and tensional forces acting along the trend of the former Devils River Uplift during uplift of the Edwards Plateau 67

29 Development of caves as related to local baselevels and to incision of the Rio Grande in the vicinity of Langtry 68
30 Schematic profile of bathymetric development of caves in the Langtry area 69
31 Location map for the Sonora area 72
32 Valley of Ice in Caverns of Sonora 72
33 Caverns of Sonora in relation to topography and caprock of the Segovia Formation 73
34 Map of Caverns of Sonora 74
35 Felton Cave in relation to topography and caprock of the Segovia Formation 76
36 Map of Felton Cave 77
37 Cross-section of Danger Pit and adjacent passage, Felton Cave 78
38 Views of Danger Pit and adjacent passage in Felton Cave 78
39 Silky Cave in relation to topography and caprock of the Segovia Formation 79
40 Map of Silky Cave 80
41 Generalized geologic section of the Sonora area showing stratigraphy and lithology 81
42 Rose-diagrams of linear segments of passages in Caverns of Sonora, Felton Cave, and Silky Cave 81
43 Generalized dip-section in Sonora area depicting sequence of development of caves 82
44 Evolution of a coralloid stalagmite 85
45 Geologic map of the Menard area 88
46 Crevice Passage in Powell Cave 89
47 Crevice Passage and overlying passages of the maze near the Big Bend in Powell Cave 89
48 Map of Powell Cave in the vicinity of the entrance 90
49 Typical passage of the maze in the vicinity of the entrance to Powell Cave 90
50 Dip-section along axis of valley of Dry Creek 91
51 Geologic map of the Bend area, San Saba and Lampasas Counties 94
52 Map of Gorman Cave 96
53 Main entrance to Gorman Cave 97
54 Main passage in Gorman Cave 97
55 Bank of sediment in Gorman Cave near the Squeeze 98
56 Bank of sediment in Gorman Cave near Separation Lake 98
57 View northeast from top of bluff directly over Gorman Cave 98
58 Terraces and bluff of limestone across Colorado River from Gorman Cave 98
59 Map of Lemons Ranch Cave 99
60 Map of El Sotanito Grande 100
61 Map of Yates Goat Cave 101
62 Map of Barnes Bat Cave 101
63 Map of Clark's Branch Well Cave 102
64 Detailed map of Area 2 of Figure 51 103
65 Maps of caves of the Gorman Fissure System 104
66 Entrance to Toad Frog Falling Floor Fissure, Gorman Fissure System 105
67 Map of Canyon Wren Cave 105
68 View northeast from top of deposit of travertine near Gorman Falls 105
69 Detailed map of Area 1 of Figure 51 106
70 Map of Soot Cave 107
71 Map of Hole in the Wall 108
72 Map of Skull Cave and Climbup Cavelet 109
73 Topographic map of the McLarrin Fissure Karst 110
74 Geologic section B-B' of the McLarrin Fissure Karst 111
75 Map of Copperhead Cave 112
76 Map of El Sotano Lodoso 113
77 Map of Gorman Sink 112
78 Map of Debris Cave 114
79 Map of Wedge Cave 115
80 Map of Blue Haw Cave 114
81 Map of Gorman Falls Cave 116

82 Map of Gormlette Waterfall Cave 117
83 Offset Fissure, McLarrin Fissure Karst 117
84 Crack Pot, McLarrin Fissure Karst. 117
85 Maps of caves of the McLarrin Fissure Karst 118
86 Meander Tube Fissure, McLarrin Fissure Karst 119
87 Dagger Cave, McLarrin Fissure Karst 119
88 Correlation of rocks of the Ellenburger Group in the Llano Region 120
89 Geologic section A-A' of the northwest part of the Bend area 122
90 Geologic section C-C' of the northwest part of the Bend area 123
91 Geologic section D-D' of the southeast part of the Bend area 124
92 Variations in physical and chemical properties of the Ellenburger Group and relation to development of caves 125
93 Rose-diagrams showing orientations of linear segments of passages in caves, mapped faults, and mapped photolineaments 126
94 Photolineaments of the Bend area 128
95 Pedestal in Gorman Cave 129
96 Solutional sculpture in Gorman Cave 129
97 Geologic map of northeastern Llano Uplift 130
98 Generalized block-diagram of the McLarrin Fissure Karst 131
99 Origin of fissure-like caves 132
100 Advanced stage of the McLarrin Fissure Karst 133
101 Recently slumped fragment of clint over Snail Fissure, McLarrin Fissure Karst 133
102 Deposit of travertine resting unconformably on Gorman Formation near Gorman Falls 134
103 Origin of caves in travertine 134
104 Geologic map of the Burnet area, Burnet and Llano Counties 138
105 Topographic and geologic map of the vicinity of Longhorn Cavern 139
106 Schematic diagram explaining origin of Backbone Ridge and Longhorn Cavern 140
107 Map of Longhorn Cavern. 141
108 Structural contour-map of the contact between the Gorman and Honeycut Formations in the vicinity of Longhorn Cavern 142
109 Geologic cross-section along plunge of synclinal axis at Longhorn Cavern 143
110 Cathedral Room in Longhorn Cavern 144
111 Generalized block-diagram of the vicinity of Longhorn Cavern 145
112 Rose-diagrams for the Burnet area 146
113 Schematic summary of major trends of fractures, regional stresses, and structural elements in the Burnet area 147
114 Generalized schematic block-diagram showing origin of Longhorn Cavern 147
115 Geologic map of the New Braunfels area, Comal and Bexar Counties 150
116 Map of the Natural Bridge Caverns System 152
117 Stereoscopic photograph of the Mount of the Landlords in the Castle of the White Giants, Natural Bridge Caverns 151
118 The Natural Bridge spanning the collapsed doline at the entrance to Natural Bridge Caverns . 151
119 Correlation of horizons of passages within Natural Bridge Caverns System, Bracken Bat Cave, and Double Decker Cave 153
120 Map of Bracken Bat Cave 154
121 Map of Double Decker Cave 155
122 Solutional cavity in the western wall of Pluto's Anteroom, Natural Bridge Caverns 157
123 Solutional cavity in the western wall of Emerald Lake Passage, Natural Bridge Caverns 157
124 Exposure of porous limestone along the eastern wall of Castle of the White Giants, Natural Bridge Caverns 157
125 Highly porous beds at the northern end of Emerald Lake Passage, Natural Bridge Caverns . . 157
126 Chapel Hall in Natural Bridge Caverns 157
127 Structural control in Natural Bridge Caverns 156
128 Dripstone formed by seepage along a bedding-plane parting, Pluto's Anteroom, Natural Bridge Caverns 158

129 En echelon faults in Fault Room of South Caverns, Natural Bridge Caverns System 158
130 Flowstone on southwestern wall of Double Decker Cave 159
131 Plan and profile of the South Caverns, Natural Bridge Caverns System, showing faults 159
132 Fractures mapped from remotely sensed imagery, New Braunfels area. 160
133 Rose-diagrams of linear segments of passages in caves and of photolineaments from aerial photographs 161
134 Breakout-dome in Hall of the Mountain Kings, Natural Bridge Caverns 163
135 Schematic diagram illustrating the evolution of breakout-domes in the Natural Bridge Caverns System 162
136 Stereoscopic photograph of redissolved formations, Hall of the Mountain Kings, Natural Bridge Caverns 164
137 Geologic map of the vicinity of Airman's Cave and Cave X, Travis County 167
138 Map of Airman's Cave, Travis County 166
139 Typical passage in Airman's Cave 167
140 Map of Cave X, Travis County 168
141 Map of Bee Creek Cave, Travis County 169
142 Map of Dead Dog Cave No. 2, Travis County 170
143 Map of Swimmin' Hole, Travis County 171
144 Map of Wonder Cave, Hays County 172
145 Map of Ezell's Cave, Hays County 173
146 Map of Wimberley Bat Cave, Hays County 174
147 Profile of Jacobs Well, Hays County 175
148 Divers at the entrance to Jacobs Well 176
149 Rose-diagrams of linear segments of passages in Airman's Cave and Cave X 176
150 Geologic map of the Georgetown area, Williamson County 180
151 Geologic sectio A-A' through the Georgetown area, showing position of the cavernous zone . . 181
152 Map of Inner Space Cavern 182
153 Origin of Inner Space Cavern 184
154 Rose-diagram showing orientations of linear segments of passages in Inner Space Cavern . . 185
155 Schematic illustration of extensional tectonism in the Georgetown area. 185
156 Topographic map of Enchanted Rock and vicinity, Llano and Gillespie Counties. 188
157 Stereoscopic aerial photographs of Enchanted Rock 189
158 Geologic map of Enchanted Rock and vicinity 190
159 Oblique aerial photograph of the northwestern side of Enchanted Rock 191
160 Gnammas near the summit of Enchanted Rock that contain water from recent rain 191
161 Gnammas developed along a joint on Enchanted Rock 192
162 Origin of gnammas 192
163 Stains indicating seepage of water emerging from sheeting joints on Enchanted Rock 192
164 Tafoni beneath exfoliated sheets on Enchanted Rock. 193
165 Stereoscopic view of exfoliated sheets on Enchanted Rock that have been pushed upward into A-tents 193
166 Map of Enchanted Rock Cave 194
167 Diagrammatic summary of the inferred evolution of Enchanted Rock Cave 195
168 Stereoscopic view of interior of Enchanted Rock Cave at the southwestern end of the lower level of the main passage 196
169 Panoramic view of large amphitheatre-like gnamma below summit of Enchanted Rock 196
170 Surveyed cross-section through Enchanted Rock Cave 197
171 Suggested areas for further detailed investigations 208
172 Number of reported caves of the Edwards Plateau and all of Texas, 1944 to 1979. 234
173 Standard symbols used on maps of caves 239
174 Method of extending lines of drainage on topographic maps 245
175 Line-intersection method of measuring drainage density 245
176 Relationship between drainage density and line-intersections 246
177 Maps of drainage density of two areas within the southeastern Edwards Plateau 247

LIST OF TABLES

1 Counties of the Edwards Plateau 27
2 Soil Surveys of the Edwards Plateau 30
3 Dominant Characteristics of Associations of Soil 31
4 Reports of the Texas Speleological Survey 33
5 Reports of Groundwater-Resources by County 34
6 Hydrogeologic Characteristics of Subdivisions of Aquifers in Bexar County 45
7 Karstic Regions of Texas 56
8 Selected Areas for Detailed Investigations 56
9 Caves of the Langtry Area 59
10 Caves of the Bend Area 95
11 Types of Passages in Inner Space Cavern. 183
12 Suggested Areas for Further Detailed Investigations 206
13 Mesozoic and Cenozoic Rocks of Central Texas 215
14 Record of Reported Caves of the Edwards Plateau by County 233
15 Significant Caves of the Edwards Plateau 235
16 Long Caves of Texas 236
17 Deep Caves of Texas 237
18 System of Grading the Accuracy of Survey-Lines 240
19 Classification of Detail in a Survey 240
20 Imagery Used in Mapping Fractures, Bat Cave Quadrangle, Comal and Bexar Counties, Texas 242
21 Evaluation of Imagery Used in Mapping Fractures 242

LIST OF PLATES

In the original typewritten version of this dissertation, the plates were folded sheets in a pocket in the back. In this reprint, they are oversized pages appended to the main file, numbered along with the rest of the text.

1 Distribution of significant caves on the Edwards Plateau, central Texas, and selected areas of study 275
2 Surficial drainage on the Edwards Plateau, central Texas 276
3 Geologic map of the Edwards Plateau, central Texas 277
4 Geologic column of Precambrian and Paleozoic rocks of the Llano Basin, central Texas 278
5 Generalized stratigraphic correlation-chart of Cretaceous units of the Edwards Plateau, central Texas. 279
6 Map of the Powell Cave System, Menard County, Texas 280

PREFACE

More than thirty-five years have passed since the publication of "Caves of Texas" (Volume 10 of the *National Speleological Society Bulletin*). In the intervening years the number of known caves in Texas has increased ten-fold through the efforts of amateur explorers. In the same period the United States Geological Survey, the Texas Department of Water Resources and its predecessors, the Bureau of Economic Geology at The University of Texas, and numerous individual researchers accelerated geological research on the Edwards Plateau, the principal region of karst in the state. Because these ancillary studies had been well underway by 1975, it was appropriate to begin a systematic, regional study on the development of caves and karst on the Plateau. I was inspired to make such an investigation after visiting many caves in the region in 1973 and 1974 while living in Houston. Having just completed a similar study on the Helderberg Plateau in east-central New York for a master's thesis, I was interested in working in a different geologic, hydrologic, and climatic setting. I had little appreciation for the scale of areas involved, let alone the scale of the scope of the project. As with many other things, a plateau in Texas is larger than one almost anywhere else.

Proposals submitted for monetary support to the National Speleological Society and the Cave Research Foundation outlined work to be accomplished in twelve selected carbonate areas within the bounds of the Edwards Plateau. In the interest of completing the dissertation, eight of these areas have been retained herein (Chapters 5 through 12). Fieldwork was largely completed in the remaining four areas (Carta Valley, Uvalde, Sabinal, and Boerne areas; see Chapter 15) and the analyses and publication of results on these should be forthcoming at later dates.

Fieldwork was carried out from 1974 to 1979. Examination, synthesis, and interpretation of the data progressed from 1977 to date. Some results were published as they became available, and these are listed in the References Cited under my authorship. Others are to follow. Investigations in the Bend, New Braunfels, Boerne, and Enchanted Rock areas were in part accomplished during contractural assignments with the Texas Natural Areas Survey, Lyndon B. Johnson School of Public Affairs, The University of Texas at Austin, from 1977 to 1978 and interpretive consultation performed for Natural Bridge Caverns in 1978.

One of the most tedious aspects of this study has been the assimilation of the existing geologic and speleologic literature into the observations, interpretations, and conclusions. The size of the area of study has made the number of consulted references inordinately large. I opted to keep all references intact in the final version of the dissertation. Thereby, the text and list of references also serve as a nearly complete synthesis of the pertinent literature. It is anticipated that this document will have added value as a basic reference for future investigators.

Many of the cited references may be obscure to the reader. Nearly all of the works dealing with Texas are available in the holdings of the libraries at The University of Texas at Austin. The speleological literature, particularly that published by speleological organizations, may be consulted at the library of the National Speleological Society, housed at its headquarters (Cave Avenue, Huntsville, Alabama).

The sources of all illustrations compiled from previously published sources are acknowledged in the captions. All photographs were taken by me, many with the assistance of my wife, Karen.

A major undertaking such as this is not accomplished without considerable assistance from numerous individuals and organizations. Dr. Victor R. Baker, currently at The University of Arizona, served as supervisor of the dissertation committee which included as members Drs. Robert L. Folk, Keith Young, and Robert K. Holz at The University of Texas at Austin and Dr. Arthur N. Palmer at the State University of New York, College at Oneonta. I am indebted to each for guidance, encouragement, and criticism during this study. Their contributed comments and suggestions have been incorporated herein. The writing has certainly improved from the meticulous editing by some members of the committee. It is difficult for a graduate student to negotiate the complexities of a large department at a large university without the assistance of the Graduate Advisor's office. I thank Dr. Allan Scott, Mary Gaddis, and Ann Page for their concern and help over the years, especially while I was enrolled *in absentia*. Kathleen Rader served as student-editor of the dissertation.

Speleological studies are impossible without the help of cavers, the enthusiasts of all things subterranean. Among those who often helpedin the gathering of data in (and under) the field are Roger Bartholomew, Ralph Batsche, Tom Byrd, Glenn and Ruth Darilek, John Graves, Scott Harden, Keith Heuss, Jan and Orion Knox, Logan McNatt, Richard Montgomery, James Pierce-Jones, Wayne W. Russell, Jr., and J. Michael Walsh. Others contributed significantly by making available maps of caves and similar speleological data or by offering their expertise during numerous discussions. They include Dr. William R. Elliott, Ronald G. Fieseler, James R. Reddell, and A. Richard Smith of the Texas Speleological Survey and Dr. Paul Boyer, Roy Jameson, David W McKenzie, Patty Mothes, William H. Russell, and George Veni.

Numerous ranchers and other landowners graciously allowed me and my companions to cross their properties and enter their caves. I especially thank Clara and Harry Heidemann of Natural Bridge Caverns, Gladys and John Bridges of Cascade Caverns, and Jolene and Eugene Ebell of Cave Without a Name for allowing me to study their caves repeatedly, even as tours were being conducted through them. My family and I made many friends among the ranchers of the Edwards Plateau during the fieldwork. Lack of space precludes listing them all, but we thank them collectively for their hospitality and help.

Many individuals from institutions concerned with studies on the Edwards Plateau provided geologic and hydrologic data. Among those at The University of Texas at Austin are: Drs. Virgil E. Barnes, E.G. Wermund, Thomas G. Gustavson, and Charles M. Woodruff, Jr. of the Bureau of Economic Geology, Donald Kennard and Dr. John Hamilton of the Texas Natural Areas Survey, Martin Smith and Chestalene Pintozzi of the Geology Library, and Dr. Ernest L. Lundelius of the Department of Geological Sciences. Drs. Glenn Longley and Albert E. Ogden of the Edwards Aquifer Research and Data Center at Southwest Texas State University forwarded results of some of their studies, as did Dr. William Back and Robert W. Maclay of the United States Geological Survey and the publication distribution office at the Texas Department of Water Resources. Drs. Thomas W. Grimshaw of Radian Corporation and L. Jan Turk of Turk, Kehle and Associates in Austin provided advice and data during the initial phase of the investigations. Fellow graduate students Robert A. Skolasky and Raymond C. Leonard shared many of their findings regarding the geomorphology and structure of the Edwards Plateau.

I was fortunate to have considerable financial assistance during the periods of research and writing. Support was provided in part by a fellowship-grant from the Cave Research Foundation, the 1977 Ralph Stone Research Award of the National Speleological Society, and two Ronald K. DeFord Field Scholarships from the Geology Foundation, Department of Geological Sciences at The University of Texas at Austin. I thank the Bureau of Economic Geology and Department of Geological Sciences for research and teaching assistantships respectively while in residence at the university and the Texas Natural Areas Survey, Radian Corporation, and Natural Bridge Caverns for opportunities of employment and consulting. In similar light, I extend my appreciation for the support of the faculties in geology and of the deans at Murray State University in Kentucky and The University of Connecticut during the difficult period of maintaining progress on the dissertation while attending to professorial duties.

The mechanics of producing a lengthy and complex manuscript are not easily accomplished without technical assistance. A.K. and Ruth Hartley contributed many hours reprographically converting the line drawings to appropriate dimensions. The Department of University Computer Systems and the Department of Geology and Geophysics at The University of Connecticut provided the programs and facilities for word-processing.

Physical speleology, as most other disciplines of geology, is best learned in the field and under the tutelage of those at the forefront of research. My mentors over the last 18 years, in the course of becoming a speleologist, include many individuals in the United States and abroad. However, my perspective has been especially tempered by Dr. Derek Ford, Dr. James F. Quinlan, Dr. Arthur N. Palmer and his wife, Margaret, and Dr. William B. White. Not only have I gained considerably from their published papers, but I have benefited from their expertise at numerous meetings, symposia, and fieldtrips to karstic regions throughout this country. To them and to many others of our avocation who are collectively called "Friends of the Karst" I express my thanks for freely sharing their knowledge and for many hours of companionship both above and below ground. I also thank Drs. Alfred Bogli of Switzerland, Petar Milanovic of Yugoslavia, and Paul Williams of New Zealand for their comradery and expertise in the field in Texas and Kentucky. Dr. C.R. Twidale of Australia provided considerable insight on geomorphic processes during our visit to Enchanted Rock and read and commented on Chapter 13.

Finally, I owe the greatest gratitude to my family. My wife Karen has invested the same amount of time in this endeavor as I have, serving as co-investigator, field-partner, technical assistant, draftsman, typist, and proofreader. Most importantly, though, she has been my companion who, during the dismal periods of overwhelming obligations and workload, provided the comfort, encouragement, and love to see this project through to its end, especially when the end was nowhere in sight. Our son, "Kass", born at the onset of this research has lived with "Daddy's Book" for all of his eight years. Unknowingly, he has provided more than his share of encouragement, cheerfulness, and joy. My parents, Ernst, Sr. and Erika, have always been highly supportive of my education, in a way that only parents can be. I thank them all.

PART I

GENERAL INTRODUCTION AND PHYSICAL SETTING

1
GENERAL INTRODUCTION

PURPOSE

Statement of Problem

The purposes of this study are (1) to ascertain the physical characteristics of surficial and subsurficial karstic landforms of the Edwards Plateau, (2) to determine to what extent particular geomorphic and hydrogeologic factors or processes have governed the development of the Edwards Plateau karst, and (3) to relate the genesis of karstic landforms spatially and chronologically to the geomorphic evolution of the region. The Edwards Plateau of central Texas was chosen as the area of study because it is one of the largest continuous karstic regions in the United States and because it is diverse in its geologic, hydrologic, climatic, topographic, and speleogenetic settings.

Needs for This Study

The Edwards Plateau covers an area approximately 430 by 240 km (Figure 1). The karstic terrain of the plateau is comparable in areal extent to the largest karstic regions of the United States, including those of the Appalachian Plateau and Valley-and-Ridge Provinces, central Kentucky and southern Indiana, the Ozark Plateau of Missouri and Arkansas, and the Pecos Valley of New Mexico (Davies, 1970). Although other major regions of karst in the U.S. have been extensively studied, karstic landforms of the Edwards Plateau or the processes which produced them have received little systematic study prior to this investigation. No regional synthesis of karst development, nor comprehensive investigations of the geomorphic evolution of the Edwards Plateau exist. An understanding of the evolution of the Edwards Plateau karst will aid in interpreting the geomorphic history of central Texas and perhaps provide additional data on Quaternary climatic changes in the region.

Knowledge of both active and ancient karstic processes in the Edwards Plateau is fundamental to meeting future economic and environmental needs, particularly along the eastern and southern margins of the Plateau where population is expanding rapidly (Muller and Price, 1979). The information derived in this study may be applied to problems of land-use capability, supply and quality of water, and disposal of waste.

SCOPE OF PRESENT STUDY

Summary of Concept

The Nature of Karst

Definition of Karst. The German word *karst* is derived from the Slovene word *kras* or *krs* and from the Italia word *carso*, meaning crag, stone, bare stony ground, or bleak waterless place (Jennings, 1971; Sweeting, 1973; Monroe, 1970). Karst, or Kras, is also the geographical name of the region of limestone in western Slovenia (Yugoslavia) "between the Gulf of Trieste in the west, the flysch (non-limestone) valley of the Vipava in the north, the basin of the Pikva in the east, and the valley of the Notranjska Reka in the southeast" (Sweeting, 1973, p. 1).

Karst is an international term used in geology, geography, and hydrology. Its exact definition is imprecise, and the term has been used with great variability in meaning by researchers throughout the world. There is no generally accepted definition. Appendix A presents several definitions of karst from recent glossaries and shows the diversity in usage of the term.

Quinlan (1978) compiled, translated, and synthesized a significant part of the world literature on karst. He introduced a comprehensive classification for types of karst based on cover, lithology, climate, geologic structure, physiography, hydrology, modification during or after karstification, and dominant landforms (Quinlan, 1967a,b, 1978). He suggested the following definitions:

Karst is ". . . an aggregate of the characteristic landforms and subsurface features produced primarily as a result of removal of minerals by solution within a rock or sediment. Solution is usually accompanied by other processes—corrasion, precipitation, transportation, sedimentation, subsidence, collapse, and eluviation into caves." (Quinlan, 1978).

"Pseudokarst consists of karst-like features, usually in non-carbonate and non-evaporite rocks, that are not produced by the karst processes mentioned above. Pseudokarst exhibits the forms (and in places even the hydrologic flow-patterns) of karst." (Quinlan, 1978).

I have chosen to use Quinlan's definitions of karst and pseudokarst in this study because they draw together several of the older interpretations. These definitions are sufficiently broad to encompass all relevant phenomena encountered in my investigations of the Edwards Plateau and other regions of karst.

*Terminology of Karst.*Research on karst is plagued by a large and complex vocabulary. Many words applied to karstic landforms and processes have evolved independently within individual countries or languages, resulting in a profusion of terms with an inordinate amount of duplication. Recently published glossaries of karstic terminology have lessened much of the confusion and have identified equivalent terms in various languages (Addison, 1973; Fenelon, 1968; Food and Agriculture Organization of the United Nations (FAO), 1972; Gams, 1973; Monroe, 1970; Paloc, 1975; Trimmel, 1965; see also Quinlan, 1978, table 1 for additional references to glossaries). The reader not familiar with terms used in this study should consult these references, in particular the glossary of Monroe (1970).

Geomorphology and Hydrogeology of Karstic Terranes

Any complete study of karst requires an interdisciplinary approach which includes (1) the description and analysis of landforms, (2) the study of the host rocks, emphasizing porosity, permeability, and solubility, (3) the analysis of the chemistry of karstic waters, and (4) the investigation of the hydrogeologic setting of the karst and its dynamics (Sweeting, 1973). Implicit in these major efforts are studies of several other factors including (1) physical and geomorphic processes such as weathering, denudation, corrasion, dissolution, subsidence, and collapse, (2) topographic evolution, (3) geologic structure, (4) climatic history, and (5) rates and durations of all processes which produce the karst.

Comprehensive reviews of the geomorphology and hydrogeology of karst have recently been completed by Jennings (1971), Sweeting (1973, 1981), Bleahu (1974), Zotl (1974), Ford and Cullingford (1976), Jakucs (1977), and Bogli (1980). Davies (1966, 1968), Quinlan (1968, 1978), LaMoreaux and others (1975a,b), Powell (1975a), and Shaw (1979) have summarized the historical development of the geomorphology and hydrogeology of karst. Both Quinlan (1978) and Shaw (1979) have cited many additional works, particularly in foreign languages, that discuss the development of karst. Early American papers on karst have been generally few in comparison with the vast European literature of the same vintage (Kastning, 1981a).

The present study stresses geomorphic and hydrogeologic influences on the evolution of karst of the Edwards Plateau. The geochemistry of groundwater, although an important consideration in karstification, is given less emphasis within the scope of this investigation.

Speleogenesis

Speleogenesis is the process of formation of caves. Cave developed by dissolution of rock are karstic landforms. Because removal of dissolved material necessitates movement of groundwater through permeable rock, speleogenesis is directly related to the hydrogeologic setting. Development of caves is a dynamic process, spatially and temporally related to the geologic setting and to various driving forces.

The origins of particular caves may vary from region to region or even from cave to cave, depending on which of many individual processes and controls dominate during the formation of caves. Often, the development of any single cave must be considered as complex interactions among numerous physical

variables and dynamic processes.

The origin of caves by dissolution has been the subject of much controversy since the early 1930's. At first, several universal theories were advanced to explain the origin of many or all solutional caves. The most notable of the American theories were advanced by Swinnerton (1929, 1932, 1935, 1942, 1948), Davis (1930), Piper (1932), Gardner (1935), Malott (1932, 1937), Rhoades and Sinacori (1941), Bretz (1942, 1953, 1954), Davies (1957, 1960), and Woodward (1961). Reviews and discussions of these and others are found in White (1959), Moore (1960), Halliday (1960), Warwick (1962), DeSaussure (1963), Waltham (1971), Kastning (1975, 1981a), Powell (1975a), and Sweeting (1981). Subsequent realization that certain caves displayed features indicative of specific modes of origin prompted geomorphologists to classify caves in genetic terms based on morphological characteristics. More recently, researchers have recognized that many types of caves are actually members of a continuum. The particular type that forms is determined by the degree to which controlling variables have dominated during speleogenesis. Some recent American investigations of the influences of specific factors on speleogenesis include those of Deike (1960), White (1960, 1966), Howard (1963, 1964a,b), Brod (1964), Thrailkill (1964, 1968), Bedinger (1966), Palmer (1969, 1971, 1972, 1975, 1977, 1981a), Rauch and White (1970), Ford (1971a,b,c, 1977a,b, 1980), Rauch (1972), Miotke and Palmer (1972), Dreiss (1974a,b, in press), Palmer (1976), and Ford and Ewers (1978).

Today caves are considered as integral components of karstic systems rather than as isolated landforms that develop independently of one another (Ede, 1975). Modern speleology focuses on karstic aquifers and treats caves as voids, channels, and conduits that give the aquifer discrete zones of porosity and permeability (Stringfield and LeGrand, 1969, 1979; Legrand and Stringfield, 1973a). This vantage allows a conceptual classification of karstic aquifers based upon both hydrogeologic setting (i.e. the factors, controls, and influences of the "classical" theories of speleogenesis) and the hydrodynamics (White, 1969, 1977). Moreover, the unified approach permits characterization of the system of flow by comparing hydrographs of discharge at springs to perturbations in recharge (LeGrand and Stringfield, 1973b). Such variations in discharge have been analyzed for several aquifers (Ashton, 1966; Ford, 1967; Wilcock, 1968; Brown and others, 1969; White and White, 1974; Knisel, 1972; Brown, 1972a, 1973; Chambers, 1973; Feder, 1973; Hess and White, 1974; Klemt and others, 1975, 1979; Dreiss, 1979, 1982, 1983). The techniques have been extended to include fluctuations in the chemistry of karstic groundwater in response to storms and seasonal conditions (White and Stellmack, 1965; Shuster and White, 1971, 1972; Jacobson and Langmuir, 1974). Additionally, the geochemistry of karstic waters is often useful in determining rates of denudation in terranes underlain by soluble rock (Harmon and others, 1972, 1975). Picknett and others (1976) review the principles and applications of the chemistry of karstic waters.

Major Emphases of Present Study

A complete understanding of the development of karst of a particular area necessitates consideration of all of the aforementioned variables, processes, and controls. However, during the course of this investigation several aspects of the problem were established as fundamental to the Edwards Plateau. Moreover, in the interest of limiting the scope so that the study may be completed in due course, this research emphasizes the following factors concerning the development of karst.

Lithostratigraphy

The carbonate strata of the Edwards Plateau consist of nearly horizontal beds of generally wide lateral extent and relatively uniform character. However, the lithic character varies considerably throughout vertical geologic sections at most specific sites. Most caves are conformable with stratification, and they occupy some beds but are excluded from others. The causal relationship between the lithic character of the rock and the excavation of caves along particular horizons is fundamental in ascertaining the vertical position of caves

Structure

Cursory examination of caves of the Edwards Plateau in plan-view indicates that most are in part comprised of linear segments. Passages are typically angulate to some degree, suggesting that they are aligned along fractures. A major emphasis of this study is to determine in detail how structure, including attitude, folds, faults, and joints have guided the flow of groundwater. Attempts are made, wherever possible, to relate systematic structures to the tectonic history of central Texas. Whereas the lithostratigraphy typically contributes to the vertical positioning of speleogenesis, structure is one of the dominant factors in determining the lateral extents, orientations, and locations of passages.

Structure is also important in the analysis of mechanical stability of caves. Areas of karstic subsidence and collapse are typically fixed along weaknesses associated with the structural setting.

Hydrogeology

The flow of groundwater through carbonate rocks obeys the fundamental laws of fluid mechanics as applied to flow in pipes and channels. Unlike most other geologic media, however, rocks undergoing karstification are subject to changes in porosity and permeability through time. As subsurficial drainage evolves in response to topographic changes, conduits along flowpaths enlarge and become better integrated. This in turn changes rates of discharge and hydraulic gradients.

Relict caves are clues in the evolution of flowpaths. Many passages and chambers of the Edwards Plateau, particularly those well above the present-day piezometric surfaces, are "fossils" of formerly active schemes of drainage. Throughout this study, caves abandoned by the water that formed them are used as indicators of paleohydrogeologic conditions. Some caves close to current baselevels are still active and are used as examples of incipient speleogenesis.

Conduits and chambers that have transmitted karstic groundwater are components of regional aquifers. The overall orientation of cavernous openings is governed by the direction of migration of groundwater prevailing during speleogenesis. Caves are simply segments of systems of flow extending from areas of recharge to points of discharge of the aquifers.

Lithostratigraphic variables and structure determine the primary porosity and permeability of the carbonate rocks at the outset. However, the hydrogeologic setting in conjunction with the evolving topography governs the spatial and temporal development of integrated flowpaths and caves. This study examines the relationship of caves to the geomorphic evolution of the Edwards Plateau and to the evolution of the major aquifers.

Topographic Evolution

The surface of the Edwards Plateau has been subjected to alternating periods of inundation by transgressive seas and denudation during intervening subaerial conditions. Moreover, large areas of the plateau have undergone epeirogenic uplift or subsidence. Structural domes, basins, arches, and troughs have imparted initial topographic relief to the region. *En echelon* faulting has produced lengthy scarps extending over several parts of the plateau and these in turn have provided sufficient topographic relief to augment hydraulic gradients and discharge within the karstic aquifer.

Caves of the Edwards Plateau are typically associated with areas of relief. Baselevels to which conduits transmit water drop as nearby streams incise the surfaces of uplands. Some caves are abandoned while others at lower elevations become active. Therefore, relict caves often reflect topographic changes. Correlation of caves to present and pre-existing topography is critical in relating speleogenesis to the geomorphic evolution of the Edwards Plateau.

Climate

Because most development of caves of the Edwards Plateau occurred during the Late Pleistocene, it is necessary to consider the climatic changes that have affected it during that interval. As indicated by ample evidence in many of the caves in the area of study, the amount of water flowing through them has varied significantly. Periods of reflooding of formerly abandoned conduits are explained by changes of climate to colder, wetter conditions during times of glaciation in the northern part of the continent. Reflooding has modified many caves of the plateau.

GEOGRAPHIC SETTING

Area of Study

The area of study includes all or parts of thirty-eight counties (Table 1 and Figure 1) in central Texas. It encompasses most of the Edwards Plateau, the Llano Basin, the Balcones Escarpment, and parts of the Lampasas Cutplain and Stockton Plateau (Figure 2). The area extends 430 km east-west and 240 km north-south and has a total area of 82,420 km^2. (For comparison, this is nearly equivalent to the area of the state of South Carolina.)

The northern boundary of the area is a line corresponding to 31°15'00" north latitude. This is close to the northern limit of outcrop for karstic Cretaceous carbonate rocks across much of the Edwards Plateau. The occurrence of karstic features significantly diminishes north of this line.

The western boundary is a line corresponding to 101°45'00" west longitude. This was chosen because it is near the Pecos River, which separates the Edwards Plateau

Table 1: Counties in the Edwards Plateau

County (Total=38)	Total area[a] (km^2)	Area in study approx. (km^2)	Percent in study	Physiographic divisions[b]	Karstic regions[c]
Bandera	1981	1981	100.0	EP	D, E
Bell	2704	1967	72.8	LCP, BP	A, B
Bexar	3230	1546	47.9	EP, BP	C, D
Blanco	1862	1862	100.0	EP, LB	C, E, F
Burnet	2587	2587	100.0	LB, LCP	A, F
Comal	1469	1469	100.0	EP, BP	C
Concho[d]	2600	878	33.8	EP, LB	H
Coryell	2701	460	17.0	LCP	A, B
Crockett	7236	6384	88.2	EP	H
Edwards	5374	5374	100.0	EP	E
Falls[d]	1971	6	0.3	BP	
Gillespie	2732	2732	100.0	EP, LB	C, E, F
Guadalupe[d]	1852	163	8.8	BP	
Hays	1735	1563	90.1	EP, BP	B, C
Irion[d]	2779	1067	38.4	EP	H
Kendall	1735	1735	100.0	EP	C, E
Kerr	2852	2852	100.0	EP	E
Kimble	3300	3300	100.0	EP	E, H
Kinney	3603	3603	100.0	EP, BP	D, E
Lampasas	1880	1326	70.5	LB, LCP	A
Llano	2440	2440	100.0	LB	F, G
Mason	2422	2422	100.0	EP, LB	F, G, H
McCulloch[d]	2761	1663	60.2	EP, BP	G, H
McLennan[d]	2678	2	0.1	LCP, BP	A, B
Medina	3504	1859	53.1	EP, BP	D
Menard	2367	2367	100.0	EP	H
Mills[d]	1901	5	0.3	LB, LCP	A
Reagan[d]	2934	808	27.5	EP	H
Real	1619	1619	100.0	EP	E
San Saba	2906	2059	70.9	LB	G
Schleicher	3447	3447	100.0	EP	H
Sutton	3867	3867	100.0	EP	E, H
Terrell[d]	6185	246	4.0	SP	J
Tom Green[d]	3973	994	25.0	EP, LB	H
Travis	2629	1893	72.0	EP, LB, LCP, BP	B
Uvalde	4113	3506	85.2	EP, BP	D, E
Val Verde	8397	8330	99.2	EP, SP	E, H, J
WIlliamson	2916	2038	69.9	EP, LB, LCP, BP	B
TOTALS	117242	82420	70.3		

[a] Source: Texas Almanac (1969).

[b] Physiographic divisions shown in Figure 2. Abbreviations: BP = Blackland Prairies, EP = Edwards Plateau, LB = Llano Basin, LCP = Lampassas Cut Plain, SP = Stockton Plateau.

[c] Letters correspond to designations of karstic regions of Figure 17.

[d] Counties that have only minor caves within the area of study

from the Stockton Plateau to the west; however, this line allows the Langtry area, one of the most significant areas of karst on the Stockton Plateau, to be included in the study. To the west of this line caves and other karstic features are generally few and small. The only notable exceptions are Sorcerer's (Adams) Cave (Veni, 1980a, 1981b; Poole, 1981) and Wizard Well (Melot, 1981, 1982) in Terrell County, which were extensively explored after completion of fieldwork for this dissertation.

The southwestern boundary is a reach of the Rio Grande extending from 101°45'00" west longitude to where the river crosses 29°05'00" north latitude. The southern and eastern boundaries of the area of study were chosen to correspond closely with (1) the southernmost and easternmost exposures of Cretaceous carbonate rocks, (2) the "bad-water line" within the artesian Edwards Aquifer, (3) the Balcones Fault Zone, and (4) the southern and eastern limits of explorable caves.

Much of the Edwards Plateau is accessible by automobile or four-wheel-drive vehicles. Nearly all land is privately owned, and much of it is included in large ranches. In some areas access is restricted by difficulty in finding or contacting the owner or obtaining permission for visitation or research. Ranches are inaccessible during deer-hunting season when the land is leased to hunters or during lambing season on sheep ranches. These factors combined with greater distances between sites have often impeded fieldwork and this may in part explain why relatively little geomorphic fieldwork has been done in the karst of the Edwards Plateau in contrast to that in karst or other regions.

Principle Geographic Regions of Central Texas

Texas has been divided into geographic regions based on physiography (Hill, 1887b, 1890a, b, 1900, 1901; Joerg, 1914; Fenneman, 1914, 1916, 1923, 1928, 1931, 1946; Atwood, 1940; Fenneman and Johnson, 1946; Raisz, 1957; Thornbury, 1965; Carr, 1967; Hunt, 1974; Kier and others, 1977), climate (Thornthwaite, 1941; Carr, 1967), soils (Godfrey and others, 1973), vegetation (Joerg, 1914; Tharp, 1939, 1944; Blair, 1950), agriculture (Texas Agricultural Experimental Station, 1960), and land resources (Godfrey and others, 1973; Kier, 1974; Kier and others, 1977). By combining some or all of these physical attributes geographers have divided Texas into "natural regions" (Johnson, 1931; Chambers, 1948; Hunt, 1974; Hezlep, 1976).

Texas has also been divided into "perceptual regions" based on physical environment, political subdivision, economy and function, terms of the compass, or promotional names (Zlatkovich, 1977; Jordan, 1978a,b; Monti, 1978). Most perceptual regions are generally coincident with physiographic regions because natural features such as fault-scarps and springs have greatly influenced the settlement, development, and economy of Texas (Buckley, 1866, p. 34-35; Simonds, 1912; Tyson, 1924; Shuler, 1936; Wilie, 1940; Bybee, 1952; Brune, 1975; Darling, 1977). The area of study, which generally coincides with the perceptual regions known as the Edwards Plateau and Hill Country, is geographically distinguished from neighboring regions by virtue of any one of the following criteria: physiography, geology, climate, soil, vegetation, agriculture, and land-resources.

Physiographic Divisions

The area of study consists of the southernmost sections of the Great Plains physiographic province and includes the Edwards Plateau and Central Texas Sections and the Blackland Prairies (Figure 2). The Edwards Plateau Section west of the Pecos River is known as the Stockton Plateau, and within the area of study the Central Texas Section is divided into the Llano Basin and the Lampasas Cut Plain (The term "Edwards Plateau," first used by Hill (1892a; see Alexander, 1976, p. 106), is derived from Edwards County, which in turn is named in honor of Hayden Edwards, an early colonizer of Texas (Texas Almanac, 1961-1962, p. 49). The Edwards Plateau is the namesake of the Edwards Limestone, one of the principal karstic stratigraphic units (Moore, 1964, p. 5). "Edwards Plateau" as used henceforth in this study refers to the entire area of study as outlined in Figure 2 and includes in whole or part the following physiographic division: the Edwards Plateau, the Stockton Plateau, the Llano Basin, and the Blackland Prairie.

On a larger scale, the Edwards Plateau may be regarded as a segment of a karsted belt that extends around the Gulf of Mexico (Fish, 1977). In Mexico the belt includes the low-lying karst of the Yucatan Peninsula and the mountainous karst of high, local relief of the Sierra Madre Orientale. To the east of the Edwards Plateau the belt extends discontinuously through Mississippi, Alabama, and Florida. Areas of karst on many of the Caribbean Islands may also be included in the belt. Because the karstic landforms evolved at various times within rocks of differing ages and under disparate hydrogeomorphic conditions, the belt is merely a physiographic province of sorts, rather than a unified morphogenetic region.

Topography

The Edwards Plateau is a tableland developed on Early Cretaceous limestone. Its interior landscape is undulating to slightly rolling and is moderately dissected by drainage to the Gulf of Mexico. The southern and southeastern margins are deeply dissected along the Balcones Escarpment, which is the topographic expression of the Balcones Fault Zone. To the west the Plateau meets the Pecos Lowland through a deeply dissected landscape. To the north the Plateau grades into the High Plains, an elevated region of low relief. Where granitic and metamorphic rocks of the pre-Cretaceous crop out in Mason, Llano, San Saba, Burnet, and Gillespie Counties, the Plateau is interrupted by the Llano Basin, a lowland of gently rolling topography. The Llano Basin grades northeastwardly into the moderately dissected,

rolling topography of the Lampasas Cut Plain (Figure 2).

The maximum elevation of the Edwards Plateau at approximately 880 m above mean sealevel, occurs in the northwest corner of the area of study. In general, the Plateau surface slopes gently to the southeast, conforming to the shallow (1° to 2°) dip of the relatively flat-lying Cretaceous strata. Minimum elevation is approximately 180 m along the base of the Balcones Escarpment in the vicinity of Austin and San Antonio. Total topographic relief within the area is about 700 m.

Local relief varies throughout the Plateau. In upland areas, where the Plateau surface is only moderately dissected, relief may be 60 m or less. However, along margins of the Plateau near the Balcones Escarpment, where streams have incised deeply, relief may be as great as 150 m.

The rugged terrain in the southern and eastern areas of the Plateau is known locally as the Hill Country. Here topographic dissection is at a maximum. Landforms commonly include narrow connecting ridges or, where erosion has been greatest, residual hills standing above local erosional plains.

Drainage

The area is drained by six major rivers (Figure 3): the Rio Grande in the southwest, the Colorado and Brazos Rivers in the north and northeast, and the Nueces, San Antonio, and Guadalupe Rivers in the south and southeast. Each of these rivers flows southeast to the Gulf of Mexico. The Nueces, San Antonio, and Guadalupe Rivers have their headwaters in the Edwards Plateau; but the Rio Grande, Colorado River, and Brazos River originate either west or northwest of the plateau.

Flow of most lesser streams on the plateau and some large streams (e.g. the Pedernales, Blanco, Comal, and Sabinal Rivers) is intermittent. Drainage is mostly underground, except during intense storms when otherwise dry valleys and arroyos maintain high discharges of short duration. However, the overall pattern of drainage on the plateau (Plate 1) is not typical of patterns found in many regions of karst because moderate-sized, perennial stream sinking abruptly into swallets are locally rare or absent. Instead, most water lost to the ground in fluvial channels enters as diffuse infiltration along reaches underlain by fractured and porous rock.

Resources of Groundwater

An important aspect of this study is that the knowledge gained about the hydrogeomorphic evolution of the karst may be applied toward an understanding of the present hydrogeologic setting and toward management of groundwater-resources. Climatic conditions, geology, and physiography contribute toward a scarcity of perennial stream on the surface of the Edwards Plateau. In order to provide water for livestock, cultivation, and municipal and domestic needs, water must be stored and then made available. Much of the water used on the Edwards Plateau comes from underground reservoirs, and catchment-tanks and surficial reservoirs contribute only a small amount (Meinzer, 1923).

Three major aquifers in the area yield large quantities of water over a large area, and two minor aquifers yield either large quantities of water in small areas or relatively small quantities of water over large areas (Figures 4 and 5). The Edwards-Trinity (Plateau) Aquifers underlie most of the Edwards Plateau and extend from Pecos, Terrell, and Upton Counties in the west to the west side of the Llano Basin in the northeast and to the dissected Plateau margin to the east and south. The Edwards (Balcones Fault Zone) Aquifer underlies Kinney, Uvalde, Medina, Bexar, Comal, Hays, Travis and Williamson Counties and lies just south and east of the Balcones Escarpment. The Trinity Group Aquifer lies beneath the Lampasas Cut Plain in the northeast sector of the area f study. The Hickory and Ellenburger-San Saba Aquifers occupy much of the Llano Basin and extend slightly to the west beneath the Edwards-Trinity Aquifer. Physical characteristics of these important aquifers are discussed in Chapter 3 as part of the hydrogeologic setting. However, because these aquifers are replenished by seepage and infiltration, the lands above them must be classified as recharge zones in land-resources terms. The most

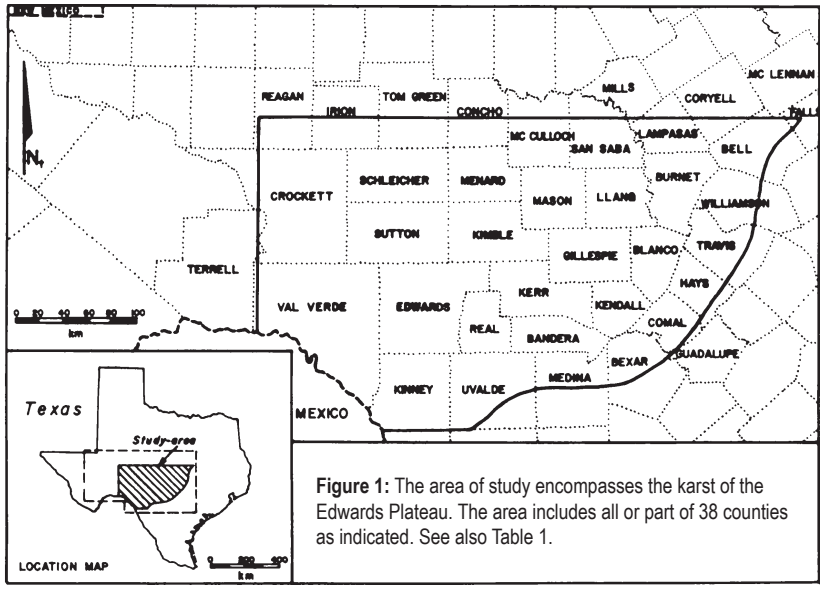


Figure 1: The area of study encompasses the karst of the Edwards Plateau. The area includes all or part of 38 counties as indicated. See also Table 1.

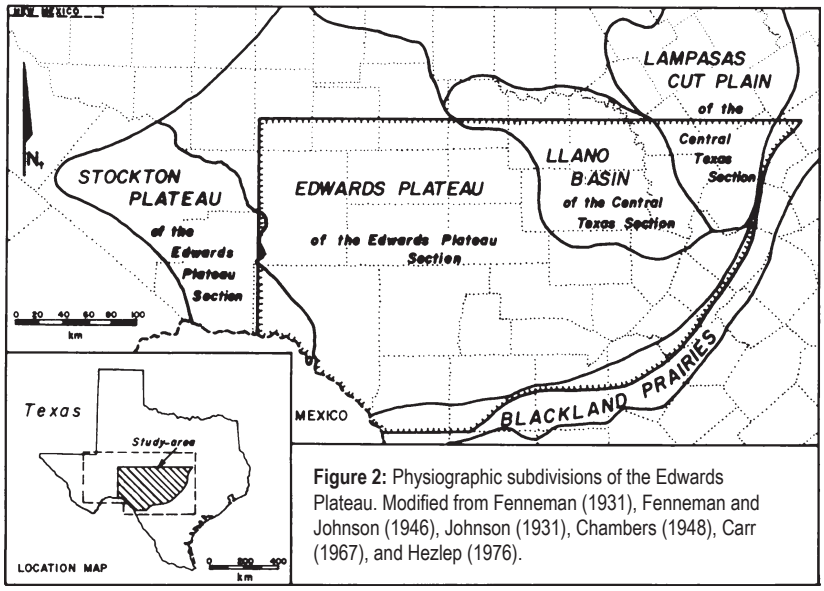


Figure 2: Physiographic subdivisions of the Edwards Plateau. Modified from Fenneman (1931), Fenneman and Johnson (1946), Johnson (1931), Chambers (1948), Carr (1967), and Hezlep (1976).

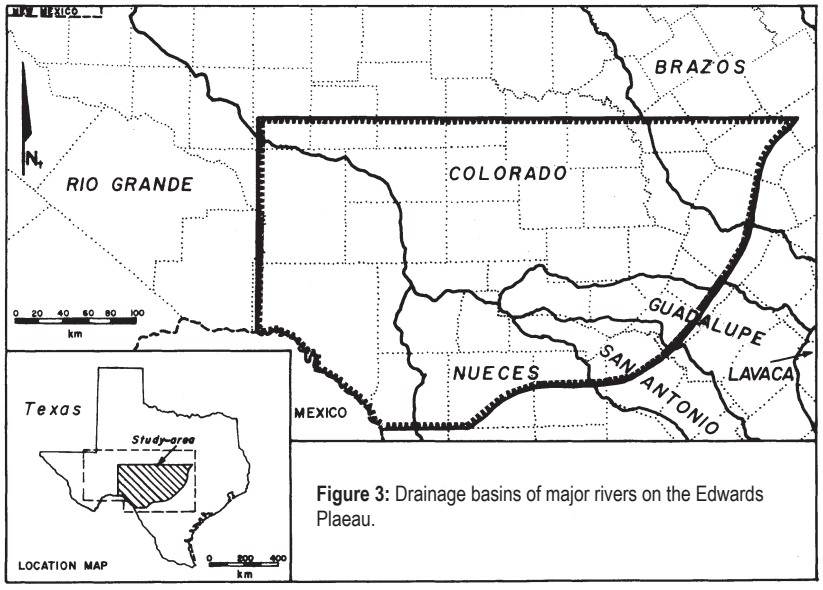


Figure 3: Drainage basins of major rivers on the Edwards Plaeau.

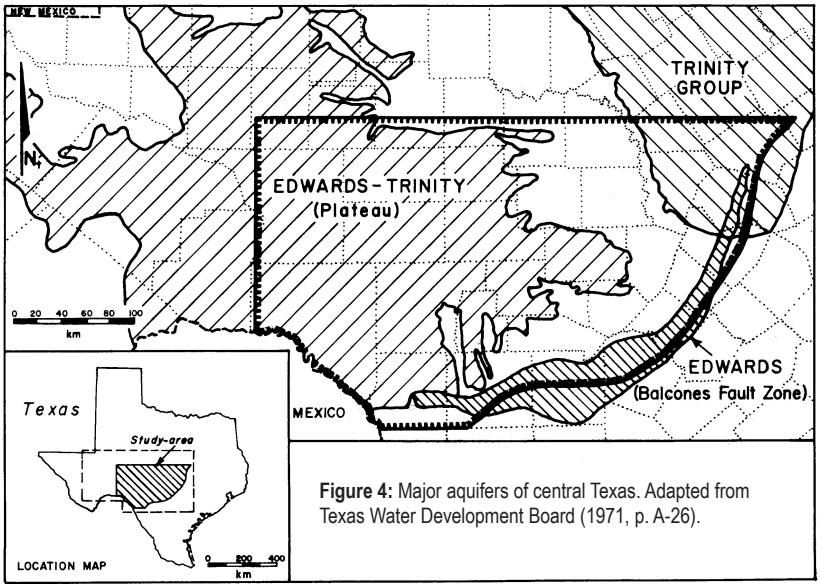


Figure 4: Major aquifers of central Texas. Adapted from Texas Water Development Board (1971, p. A-26).

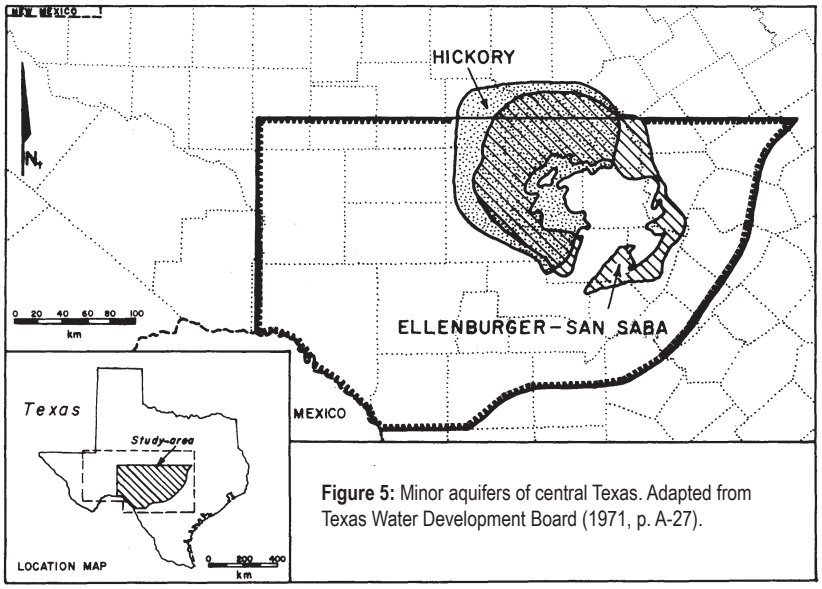


Figure 5: Minor aquifers of central Texas. Adapted from Texas Water Development Board (1971, p. A-27).

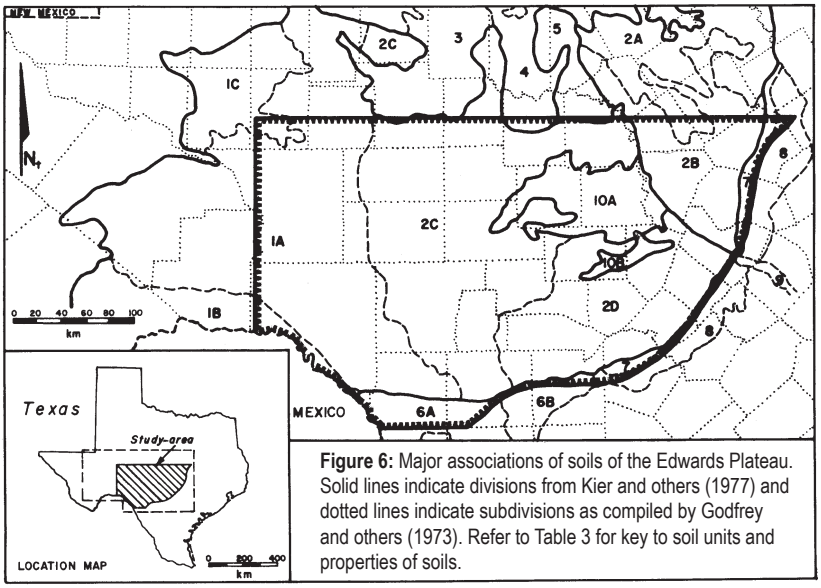


Figure 6: Major associations of soils of the Edwards Plateau. Solid lines indicate divisions from Kier and others (1977) and dotted lines indicate subdivisions as compiled by Godfrey and others (1973). Refer to Table 3 for key to soil units and properties of soils.

important area of recharge lies directly on the outcrop of Edwards Limestone within the Balcones Fault Zone. Here water that enters the bedrock recharges the Edwards Aquifer to the south and east. This aquifer is the principal supply of water for San Antonio and other cities along the fault zone from Bracketville, Kinney County to San Marcos, Hays County.

Several Large reservoirs have been constructed along major streams on the Edwards Plateau. Among them are Buchanan Lake, Inks Lake, Lake Lyndon B. Johnson, Lake Travis, and Lake Austin on the Colorado River, Canyon Lake on the Guadalupe River, Medina Lake on the Media River, and Lake Amistad on the Rio Grande. There are many additional but smaller impoundments on surface streams and construction of others, large and small, have been proposed (United States Army Corps of Engineers, 1965). Some large reservoirs on the Edwards Plateau have been designed to serve multiple purposes including flood control, generation of hydroelectric power, irrigation, and recreation. Other data on reservoirs of central Texas can be found in Plummer (1944a, 1946), Texas Water Commission (1956), McDaniels (1964), Dowell (1964), Dowell and Breeding (1967), and Dowell and Petty (1971).

Soils

Soils of the Edwards Plateau are generally thin and stony and consist of dark calcareous clay and clay loam (Kier and others, 1977). In areas where units of limestone are thin-bedded and slopes are steep, particularly in the moderately to deeply dissected karst, bare ledges of rock are common, and soils may be locally absent (Wermund, Cannon, and others, 1974; Kier and others, 1977). Soils developed on crystalline rocks in the Llano Basin are reddish-brown to brown, gravelly and stony, sandy loam. Those in the Lampasas Cut plain are dark-colored, deep to shallow, clay loam, clay, and stony, calcareous clay where they overlie limestone (Kier and others, 1977).

Several investigators have analyzed Texas soils, ascertained their physical characteristics, and classified them accordingly (Dumble, 1895; Carter, 1931, 1944; Jones and Rosamond, 1946; Texas Agricultural Experimental Station, 1960; Godfrey and others, 1973). Others have related soil development to existing, physical environments, including geology, climate, and vegetation (Chambers, 1941, 1942; Baker, 1944; Godfrey and others, 1973; Kier and others, 1977) and to resources, capability, and use of the land (Texas Agricultural Experimental Station, 1960; Godfrey and other, 1973; Kier and others, 1977).

To date, the United States Soil Conservation Service has published detailed soil surveys for 23 additional counties in the area of study, reconnaissance surveys for larger areas which include several additional counties, and surveys for smaller areas in the vicinity of San Antonio, San Marcos, and Austin (Table 2). All but four of the individual county reports in Table 2 have been completed since 1965 and contain maps of soils drawn on reproduced, serial photographs at scales of 1:20,000 or 1:40,000. Figure 6 shows the major soil associations of the Edwards Plateau as compiled from general soil maps of Godfrey and others (1973) and Kier and others (1977). Table 3 lists the dominant characteristics of these associations.

PREVIOUS INVESTIGATIONS AND REVIEW OF THE LITERATURE

The Edwards Plateau is one of the largest, yet least studied karstic regions of the United States. For example, Davies and LeGrand (1972) devote 38 pages to a review of the principal karstic regions of the United States, but only one half a page is alloted to the Edwards Plateau. This neglect is made more apparent by the known occurrence of over 1800 caves from the Edwards Plateau. Because many large areas are difficult to reach from public roads and have not been checked for caves, undoubtedly hundreds, if not thousands, of additional caves remain to be discovered and catalogued.

Even though the geomorphology of the Edwards Plateau karst has remained relatively unstudied, the basic geology and hydrogeology of the Plateau have been under continual investigation. Investigations of minerals and water-resources account for much of this activity.

Table 2: Soil Surveys of the Edwards Plateau

County or Area	Reference
Bandera County	Hensell and others (1977)
Bell County	Carter and others (1918); Huckagee and others (1977)
Bexar County	Taylor and others (1966)
Blanco County	Dittemore and Allison (1979)
Burnet County	Dittemore and Allison (1979)
Falls County	Beck and Carter (1932); Wyrick and others (1978)
Gillespie County	Allison and others (1975)
Guadalupe County	Ramsey and Bade (1977)
Kendall County	Dittemore and Allison (1979)
Kimble County	Blum (1982)
Kinney County	Newman and others (1967)
McCulloch County	Bynum and Coker (1974)
McLellan County	Templin and others (1958)
Medina County	Dittemore and Allison (1979)
Menard County	Coffee and others (1967)
San Saba County	Veatch and others (1917); Bynum and others (1982)
Schleicher County	Wiedenfeld (1980)
Sutton County	Wiedenfeld and McAndrew (1968)
Terrell County	Turner and Fox (1974)
Tom Green County	Wiedenfeld and Flores (1976)
Travis County	Werchan and others (1974)
Uvalde County	Stevens and Richmond (1976)
Williamson County	Templin and others (1936)
South-Central Texas (Reconnaissance)	Kocher and party (1915)
Southwest Texas (Reconnaissance)	Kocher and party (1912)
Trans-Pecos Area (Reconnaissance)	Carter and others (1928?)
West-Central Texas (Reconnaissance)	Carter and others (1928)
Austin Area	Mangum and Belden (1905)
San Antonio Area	Caine and Lyman (1904)
San Marcos Area	Mangum and Lyman (1906); Lowther (1972)
Waco Area	Mangum and Carr (1906)

Existing Data-Base

Substantial topographic, geologic, and hydrologic information is available on the Edwards Plateau and provides a data-base for this investigation. The area of study encompasses 526 7.5-minute U.S.G.S. topographic quadrangles. Nearly all of these have been mapped at a scale of 1:24,000 and the few that have not have been mapped at a scale of 1:63,360. Most 7.5-minute quadrangles have been published only within the last 15 years (Baskin, 1980).

Geologic maps of the Edwards Plateau are available at a scale of 1:250,000 as part of the Geologic Atlas of Texas of the Bureau of Economic Geology of the University of Texas at Austin. Nine sheets of this series cover the area of study (Barnes, 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b). Specific localities in the area of study have been mapped in greater detail, particularly in the Llano Basin and along the Balcones Fault Zone. Brown (1963) lists all geologic maps completed prior to 1962. Most large-scale geologic maps completed since 1962 are cited on sheets of the Geologic Atlas of Texas.

The U.S. Geological Survey and Texas Department of Water Resources have maintained extensive records of surficial and subsurficial water, including those of stream flow, spring flow, water levels in wells, and water quality. The U.S. Weather Service maintains records of precipitation on the Edwards Plateau. These data are used to determine responses of streams and aquifers to precipitation, including events of short duration such as storms and events of long duration such as wet periods and droughts.

Environmental, geologic maps, and photographic mosaics (scale 1:24,000) for the southern Edwards Plateau were produced in conjunction with a comprehensive, environmental, geologic analysis of the basins of Nueces, San Antonio, Guadalupe, and Lavaca Rivers of south Texas (Morton, 1974; Wermund, 1972, 1974a,b; Wermund and True, 1972; Wermund, Cannon, and others, 1974; Wermund and Woodruff, 1977). Maps showing environmental geology, physical properties, active processes, biologic assemblages, economic resources, land use, and slopes are on file with the Bureau of Economic Geology of the University of Texas and the Texas Department of Water Resources. At this writing, these maps are being readied for publication at a scale of 1:250,000 (Wermund and Gustavson, 1983). Maps and photographic mosaics showing lineaments and fractures in the southern plateau were produced during a study of regional distribution of fractures and the relationship of fractures to regional movement of groundwater, tectonics, and caves (Wermund 1973, 1976; Wermund, Cepeda, and Bell, 1974; Wermund and Cepeda, 1977a,b; Wermund and others, 1978). These materials are on file at the same agencies.

Speleological data, including descriptions, maps, and statistical information on Texas caves, are on file with the Texas Speleological Survey (housed in the Texas Memorial Museum, Austin, Texas), and several compilations have been published for the Edwards and Stockton Plateaus (Table 4, see also Widener, 1958). Data on caves of the plateau are summarized in Chapter 4 and Appendix C. The techniques employed in mapping caves are described in Appendix D.

Investigations in the Edwards Plateau

Geologic Problems

The general geology of much of the Edwards Plateau has been described in reports of reconnaissance of the U.S. Geological Survey and Texas Department of Water Resources (formerly the Texas Board of Water Engineers, Texas Water Commission, and Texas Water Development Board), especially for drainage basins and counties for which hydrogeologic studies of the Edwards and Trinity Aquifers have been underway for some time (Alexander and others, 1964; Brown and others, 1965; Cronin and others, 1963; Davis and Leggat, 1965; Maclay and Small, 1976; Mount and others, 1967; Rees and Buckner, 1980; Walker, 1979; and the county reports listed in Table 5). Substantial literature exists describing the stratigraphic relations and areal distribution of carbonate rocks of central Texas (principally Hill, 1887c, 1890c, 1891c, 1929; Adkins, 1932; Sellards, 1932a; Cloud and Barnes,

Table 3: Dominent Characteristics of Association of Soil[a]

Map unit	Soil association[b]	Setting		Mineralogical and chemical properties	Dominent vegetation	Major land uses and capabilities
		Regolith	Reflief			
1-A 1-B 1-C	Ector[o]-Conger[f]-Rock outcrop Lozier[e]-Upton[f]-Rock outcrop Reagan[e]-Conger[f]-Ector[o]	Calcareous, loamy materials over indurated limestone or weakly cemented to indurated caliche.	Steep to level.	Carbonatic; moderately alkaline and calcareous throughout.	Short and mid grasses; desert schrub; scrubby live oak (Ector).	Range; crops (irrigated - Reagan); wildlife; recreation.
2-A 2-B 2-C 2-D	Denton[o]-Purves[o]-Brackett[l] Tarrant[o]-Brackett[l]-Denton[o] Tarrant[o]-Kavett[o]-Tobosa[g] Tarrant[o]-Brackett[l]-Speck[m]	Calcareous, clayey and loamy materials over indurated limestone, interbedded limestone, and marl; calcareous, clayey outwash.	Level to hilly.	Mixed, montmorillonitic (Tarrant); carbonatic (Brackett); moderately alkaline; mostly calcareous.	Tall and mid grasses; live oak savanna; also juniper (Brackett); some mesquite trees.	Range; crops (Denton); wildlife; recreation.
3	Rowena[o]-Sagerton[n]-Mareta[o]	Clayey and loamy outwash from red beds and limestone prairies; calcareous, clayey red beds.	Level to undulating.	Mixed (appreciable montmorillonite), neutral in upper part, moderately alkaline and calcareous below.	Short and mid grasses; mesquite trees; cacti.	Crops; crops (irrigated); range; wildlife.
4	Tarrant[o]-Kavett[o]-Rowena[o]	Calcareous, loamy and clayey materials over silty and clayey red beds or limestone; loamy and clayey outwash.	Rolling to level.	Mixed; montmorillonitic (Tarrant); moderately alkaline and calcareous below.	Short and mid grasses; tall oak trees (Tarrant); mesquite trees.	Range; crops (irrigated - Rowena); wildlife.
5	Truce[o]-Owen[l]-Waurika[j]	Clayey and loamy, noncalcareous materials from shale, sandstone, and outwash.	Level to hilly.	Mixed or montmorillonitic (Waurika); acid or moderately alkaline (Owens) in upper part, moderately alkaline and calcareous below.	Mid grasses; short grasses (Owens); mesquite and post oak trees.	Range; crops (Waurika); wildlife.
6-A 6-B	Uvalde[o]-Montell[p]-Zapata[f] Knippa[o]-Castroville[o]-Atco[l]	Calcareous, clayey, and loamy marine and deltaic sediments; loamy to clayey stream alluvium and outwash sediments.	Level to undulating.	Montmorillonitic or mixed (appreciable montmorillonite); or carbonatic (Zapata); moderately alkaline and calcareous throughout.	Short and mid grasses; mesquite trees; thorny schrubs; cacti.	Range; crops; crops (irrigated); wildlife.
7	Austin[l]-Stephen[l]-Eddy[h]	Marine chalk and interbedded chalk and marl.	Gently undulating to gently rolling.	Mixed or carbonatic (Eddy); calcareous.	Tall grasses (also juniper on Eddy).	Urban; pasture; residences; crops; range.

Continued on next page.

Table 3: continued

Map unit	Soil association[b]	Setting		Mineralogical and chemical properties	Dominent vegetation	Major land uses and capabilities
		Regolith	Reflief			
8	Houston Black[p]-Heiden[q]-Austin[l]	Calcareous, clayey marine marl and chalk.	Nearly level to undulating.	Montmorillonitic or mixed (primarily montmorillonite and clay-size carbonates in the Austin); calcareous throughout.	Tall grasses.	Crops; pasture; range; urban.
9	Miller[l]-Norwood[g]-Pledger[k]	Clayey and loamy, calcareous, recent flood-plain alluvium.	Level.	Mixed (appreciable montmorillonite); high base status; moderately alkaline and calcareous to neutral in surface layer, moderately alkaline and calcareous below.	Hardwood forest; shade tolerant tall grasses.	Crops; crops (irrigated); pasture; urban; parks.
10-A 10-B	Castell[c]-Pontotoc[c]-Ligon[d] Pedernales[c]-Hensley[d]- Pontotoc[c]	Clayey and loamy, non-calcareous unconsolidated materials over granite, schist, gneiss, limestone, and sandstone, and from outwash.	Level to hilly.	Mixed; acid to mildly alkaline; calcareous in lower part (Pedernales).	Tall grasses; mid grasses; live oak motts; scrubby post oak trees; mesquite trees; shrubs; cacti.	Range; crops; wildlife; recreation; residences with community septic systems.

[a]Adapted from Godfrey and others (1973).
[b]Superscripts indicate great groups of soils according to taxonomy of Soil Survey Staff (1975) as follows:
Alfisols, including cPaleustalfs and dRhodustalfs.
Aridisols, including eCalciorthids and fPaleorthids.
Entisols, including gUdifluvents and gUstorthents.
Inceptisols, including iUstochrepts
Mollisols, including jArgialbolls, kHapludolls, lHaplustolls, mArgiustolls,nPaleustolls, and oCalciustolls
Vertisols, including pPellusterts and qChromusterts

1946a; Tucker, 1962; Winter, 1962; Lozo and Smith, 1964; Moore, 1964; Rodda and others, 1966; Young, 1967; Stricklin and others, 1971; Rose, 1972; see also Appendix B). Numerous geologic investigations of specific areas within the boundaries of the present area of study have been completed and are indexed in published bibliographies on Texas geology (Sellards, 1932b; Girard, 1959; Brown, 1963; Moore and Brown, 1972; Moore, 1976; Masterson, 1981). Many references included therein are theses and dissertations submitted primarily to universities in Texas and contain a wealth of stratigraphic information as well as geologic maps. Previous investigations pertaining to specific sites within the current area of study are cited in appropriate chapters of Part II.

Geomorphic Problems

Geomorphic studies in central Texas were few prior to 1940. Work began with studies of the geographic regions of Texas (Hill, 1887a, 1890a,b, 1891a, 1892, 1900, 1901; Tarr, 1894), the Edwards Plateau and its underground waters (Hill and Vaughan, 1898a), the topography of Texas (Hill, 1887b, 1890b, 1891b; Tarr, 1890a), and drainage in central Texas (Hill, 1889; Tarr, 1890b). Physiographic descriptions of the Nueces, Austin, Uvalde, and Llano-Burnet 30-minute quadrangles were published by the United States Geological Survey during the same period (Hill and Vaughan, 1898b, 1902; Vaughan, 1900; Paige, 1912a).

A considerable number of investigations of geomorphic problems of central Texas were initiated since 1940, but most were published within the last 15 years. The studies concentrated largely on fluvial processes and emphasized transport and deposition of material by streams flowing across the Edwards Plateau or characteristics of terraces along these rivers. Several investigators studied sediments derived from the Colorado River, which drains an area of diverse lithology that includes terrigenous, carbonate, metamorphic, and igneous rocks (Mathis, 1942, 1944; Sneed and Folk, 1958; Weber, 1968; Bradley, 1970). Others studied deposition and morphology of Pleistocene terraces along the Colorado River (Weeks, 1941, 1945a; Quinn, 1957; Urbanec, 1963; Mandel, 1978), Brazos River (Bronaugh, 1950; Stricklin, 1953, 1957, 1961), Blanco River (Koenig, 1940), and Sabinal River (Mear, 1953). Additional work on the Colorado River over the last few years has focused on its adjustments to climate and hydrologic regimen during the Quaternary (Penteado and Baker, 1974; Baker and Penteado, 1975; Baker and Penteado-Orellana, 1977, 1978; Sorenson and others, 1976; Wallis, 1976; Looney, 1977; Looney and Baker, 1977; Skolasky, 1978).

Fluvial investigations have also addressed the origin of incised meanders of Edwards Plateau streams (Blank, 1970a,b; Tinkler, 1971, 1972, 1973) and the quantitative geomorphology of small watersheds (Shepherd, 1975a,b, 1979; Hulke, 1978). There has been an active interest in the geomorphic evolution and paleohydrology of major rivers of central Texas including that of the Colorado River (Wallis, 1976), Brazos River (Epps, 1973), Leon River (Lewand, 1969), Pecos River (Leonard and Frye, 1962; Thomas, 1972; Patton and Dibble, 1978), and Rio Grande (Belcher, 1975). Woodruff (1974, 1977), Woodruff and Abbott (1979a,b), Baker and Woodruff (1974), and Canon (1974) have discussed the role of stream-piracy during evolution of drainage in the vicinity of the Balcones Escarpment. Shepherd (1975a,b, 1979) has presented evidence for stream-piracy in the Llano Basin.

Perhaps the greatest emphasis to date regarding fluvial geomorphology on

the Edwards Plateau has been on the effects of catastrophic flooding due to severe storm. Recent work includes analyses of the morphology of streams and the geomorphic response of channels to high runoff (Baker, 1974, 1977; Baker, Patton, and Shepherd, 1975; P.C. Patton, 1977; Patton and Baker, 1975, 1976, 1977; Baker and Woodruff, 1974; Patton and others, 1979; Kochel, 1980; Kochel and Baker, 1982). Baker and co-workers have formulated methods to recognize and evaluate flood hazards (Baker, 1975, 1976; Baker and Holz, 1978; Baker and others, 1974; Baker, Holz, and others, 197; Baker, Holz, and Patton 1975; Baker and Patton, 1975).

Interest in the evolution of the regional topography of Texas is being renewed (Menzer and Slaughter, 1970). A key to developing a geomorphic history for the Edwards Plateau is an understanding of the age and origin of sediments derived during denudation of the region (McFarland, 1939; Young, 1972b).

Hydrogeologic and Geochemical Problems

Most geologic studies of the Edwards Plateau have arisen from the need to study its water-resources. Various aquifers of the Plateau have been vigorously investigated, particularly those in the vicinity of the Balcones Fault Zone and just to the south and east of the zone, where the Edwards Limestone underlies Upper Cretaceous and Tertiary units. Hill and Vaughan (1898a) and Hill (1893, 1900) were the first to study the hydrogeology of the Plateau and its aquifers. Other hydrogeologic studies followed as supplies of groundwater became increasingly important to cities along the Balcones Escarpment, particularly to the San Antonio area (Livingston and others, 1936; Sayre and Bennett, 1942; Guyton and Rose, 1945; George, 1948; Petitt and George, 1956; George and others, 1962; Weinert, 1964; United States Army Corps of Engineers, 1965; Green, 1967; Hammond, 1972; E. Black, 1973; Maclay and Buckner, 1980; and the county reports of Table 5). Today, greater San Antonio, which has a population of about one million people, obtains nearly all of its water from the Edwards Aquifer (Edwards Underground Water District, 1979). The hydrogeology of the aquifer is presently under renewed study (Klemt and others, 1972, 1975, 1979; Puente, 1975; 1976; Sieh, 1975; Maclay and Small, 1976, 1979, 1983), and discharges of stream and springs, waterlevels in observation wells , and water-quality are continuously monitored as they have been since the 1930's (Lang, 1954; Garza, 1962, 1966; Rettman, 1969; Maclay and Rettman, 1972; Maclay and Rappmund, 1979; Guyton and Associates, 1979; Marquardt and Elder, 1979).

Reconnaissance investigations of groundwater-resources have been published for basins of the Brazos River (Cronin and others, 1963), Colorado River (Mount and others, 1967), Guadalupe, San Antonio, Nueces River (Alexander and others, 1964), and Rio Grande (Peckham, 1963; Leggat, 1965; Brown and others, 1965; Baker, 1965). A few recent studies have centered on the hydrochemical evolution of the Edwards Aquifer (Abbott, 1973, 1974, 1975, 1977a,b; Deike and Pearson, 1978; Woodruff and Abbott, 1979a,b; Land and Prezbindowski, 1981), the role of groundwater in the diagenesis of Cretaceous carbonates (Henningsen, 1962; Mench, 1978; Land and Mench, 1978), and isotopic studies of carbonate groundwaters of the Edwards Plateau (Kronfeld and Adams, 1971, 1974; Kronfeld, 1972, 1974; Rightmire and others, 1974; Pearson and others, 1975; Pearson and Rettman, 1975; Browning, 1977; Rye and others, 1981; Kreitler and Browning, 1983).

Management of water-resources of the Edwards-Trinity aquifers has prompted investigations of the potential of both natural and artificial recharge (Knisel, 1965; Brune, 1966; Blank and others, 1966; Green, 1967; Pool, 1973, 1977). As part of a study of the environmental geology of the southern Edwards Plateau, the Bureau of Economic Geology at the University of Texas at Austin has investigated the nature and distribution of fractures in that region with emphasis on how they influence the flow of groundwater (Wermund, 1973, 1976; Wermund, Cepeda, and Bell, 1974; Wermund and Cepeda, 1977a,b; Wermund and others, 1978). Leonard (1977a,b, 1978) extended this work to the west into Val Verde County. Several other studies have been concerned with the control of geologic structure on the flow of groundwater at specific localities (e.g. Kiersch, 1952; Kiersch and Hughes, 1952a,b; Billa, 1957; Henderson and others, 1959; Grimshaw, 1970; and Shaw, 1974, 1978; Abbott, 1977a; Kastning, 1980b, 1981b, 1983a,b, 1984).

There are thousands of springs on the Edwards Plateau, and many larger ones located along the deeply dissected margins of the plateau serve as points of discharge for the carbonate aquifers. Meinzer (1927) and Brune (1975) have catalogued major springs in the area and compiled data regarding their size, discharge, historical significance, and economic importance.

Most previous hydrogeologic studies in the area of study concern groundwater of the Cretaceous carbonate aquifers. Few published reports address the minor, carbonate aquifers. However, there has been some recent interest in the availability and quality of groundwater from the crystalline rocks in the Llano Basin (Landers, 1972; Landers and Turk, 1973).

Environmental and Engineering Problems

Areas of karst are typically prone to environmental and engineering problems in cases where and when the land becomes heavily used or developed. Major concerns include (1) transmittal of pollutants, (2) denudation of soil, (3) collapse of dolines, (4) availability and predictability of groundwater, (5) scarcity of surficial water, and (6) leakage of reservoirs and other impoundments (LeGrand, 1973, 1977).

Environmental problems of the Edwards Plateau have heretofore been limited in areas of rapid urbanization, particularly in and near cities and large towns along the Balcones Escarpment (Garner, 1972; Wermund and True, 1972; Wermund, 1972, 1974a,b; Cannon, 1974b,c; Shaw, 1974; Wermund, Cannon, and others, 1974; Wermund and Waddell, 1974; Plamondon, 1975; Woodruff, 1975, 1979; Garner and Young, 1976; Grimshaw, 1976; Wermund and Woodruff, 1977; Wermund and Gustavson, 1983). Some studies have addressed environmental implications of water-quality in these sensitive areas (George and Hastings, 1951; Lyndon B. Johnson School of Public Affairs, 1971; Garner and Shih, 1973; Evans, 1974; Turk, 1974, 1975; Browning, 1977; St. Clair, 1979; Townsend,

Table 4: Reports of the Texas Speleological Survey[a]

Issue[b]	Editor(s)	Date	Contents or Subject
	J. R. Reddell	1961a	Preliminary checklist
v. 1, #1	J. R. Reddell, W. H. Russell	1961a	Travis County
v. 1, #2	J. R. Reddell, W. H. Russell	1961b	Langtry Area (Val Verde County)
v. 1, #3	J. R. Reddell	1961b	Uvalde County
v. 1, #4	J. R. Reddell, O. Knox	1962	Bexar County
v. 1, #5	J. R. Reddell, W. H. Russell	1962	Checklist
	anonymous	1962a	Additions to checklist
	anonymous	1962b	Corrections to checklist
v. 1, #6	J. R. Reddell, J. Estes	1962	San Saba County
v. 1, #7	J. R. Reddell	1963	Val Verde County
v. 1, #8	J. R. Reddell, W. H. Russell	1963	Northwest Texas
v. 2, #1	J. R. Reddell, R. Finch	1963	Williamson County
v. 2, #2	J. R. Reddell	1964a	Comal County
v. 2, #3	R. M. Frank	1964	Vertebrate Paleontology
v. 2, #4	D. W. MacKenzie, J. R. Reddell	1964	Bell and Coryell Counties
v. 2, #5-6	J. R. Reddell, A. R. Smith	1965	Edwards County
v. 2, #7	A. R. Smith, J. R. Reddell	1965	Kinney County
v. 2, #8	J. R. Reddell, A. R. Smith	1966	Revised checklist
v. 3, #1	J. R. Reddell	1967	Medina County
v. 3, #2	C. E. Kunath, A. R. Smith	1968	Stockton Plateau
v. 3, #3	J. R. Reddell	1968	Bibliography
v. 3, #4	J. R. Reddell	1970	Lubbock County
v. 3, #5	T. Mollhagen	1970	Key to Bats of Texas
v. 3, #6	A. R. Smith, J. R. Reddell	1971	Kimble County
v. 3, #7-8	J. R. Reddell	1973	San Saba County
v. 4, #1	R. G. Fieseler, C. E. Kunath	1975	Brewster and Western Pecos Counties
v. 4, #2	J. R. Reddell, R. G. Fieseler	1977	Far West Texas

[a]These reports were published in small quantities at irregular intervals; consequently they are quite rare. Copies have been deposited in the Library of the National Speological Society and the Geology and Barker Texas History Libraries at the University of Texas at Austin. Complete citations appear in the References Cited.
[b]Issues up to and including v. 2, #1 were dittoed and v. 2, #2-8 were mimeographed. Succeeding issues were printed by photo offset. Many of these issues have been reprinted, at least in part, in the *Speleo Digest* series of the National Speleological Society.

Table 5: Reports of Groundwater Resources by County[a]

County[b]	Groundwater resources	Well and spring inventories; water levels	Other
Bandera	B6210 Reeves and Lee (1962)		
Bell		B5902 Rayner (1959a)	
Bexar	B5911 Arnow (1959) U013 Livingston (1947) W1588 Arnow (1963)	B5606 Follett (1956a) R237 Marquardt and Elder (1979)	
Blanco	R174 Follett (1973)	U014 Barnes and Cumley (1942)	
Burnet			M62-01 Mount (1962)
Comal	U059 George and others (1947) W1138 George and tohers (1952)	B5610 Follett (1956c) U058 Michal (1937)	
Crockett	B47 Inglehart (1967)	B5903 Rayner (1959b)	
Edwards	B6208 Long (1962) W1619J Long (1963)	U078 Turner and Frazier (1939)	
Gillespie		U093 Turner and Shield (1937)	M63-03 Mount (1963)
Guadalupe	R19 Shafer (1966)	B5610 Follett (1956c) U103 Altgelt and Michal (1937)	
Hays	B6004 DeCook (1960) W1612 DeCook (1963)	B5501 DeCook and Doyel (1955) B5612 Follett (1956a) U114 Barnes (1938)	
Irion		U142 George and Frazier (1941)	R145 Pool (1972b)
Kendall	R60 Reeves (1967)	U152 George and Frazier (1940a)	B5204 George and Doyel (1952)
Kerr	R102 Reeves (1969)		
Kimble	R95 Alexander and Patman (1969)		
Kinney	B6216 Bennett and Sayre (1962)	B5611 Follett (1956d) U154 Bennett and Cromack (1940)	
Mason		U182 George and Lyle (1940)	
Medina	B5601 Holt (1956) W678 Sayre (1936) W1422 Holt (1959)	B5609 Follett (1956b)	
Menard	B6519 Baker and others (1965)		
McLennan		B5902 Rayner (1959a)	
Reagan		B5903 Rayner (1959b)	B145 Muller and Couch (1972)
Real	B5803 Long (1958)		
San Saba		U243 George and Shafer (1939)	U244 George (1944)
Schleicher			R132 Muller and Couch (1971)
Sutton			R147 Muller and Pool (1972)
Terrell		B5903 Rayner (1959b)	
Tom Green	B5411 Willis (1954)	B5907 Rayner (1959c) U279 Barnes and Dalgarn (1941)	
Travis		B5612 Follett (1956e) B5708 Arnow (1957) U282 George and others (1941)	
Uvalde	B6212 Welder and Reeves (1962) W 678 Sayre (1936) W1584 Welder and Reeves (1964)	B5611 Follett (1956d)	
Val Verde	R172 Reeves and Small (1973)	B5611 Follett (1956d) U285 George and Frazier (1940b)	
Williamson		B5612 Follett (1956e) U298 Cumley and others (1942)	

[a]Numbered reportes are prefixed as follows: B- Bulletins of the Texas State Board of Water Engineers and the Texas Water Commitssion, M- Memorandum Reports of the Texas Water Commission, R - Reports of the Texas Warer Development Board and Texas Department of Water Resources, U - Unnumbered reports of the Texas State Board of Water Engineers (numbers refer to index numbers In the 1981 Publications Catalog of the Texas Department of Water Resources), W = Water-Supply Papers of the U. S. Geological Survey.
[b]Reports of groundwater-resources have not been published for the following counties: Concho, Coryell, Falls, Lampassas, Llano, McCulloch, and Mills.

1981; Kreitler and Browning, 1983).

Engineering problems related to the karst of the Edwards Plateau typically stem from the construction of dams and reservoirs (Dowell and Petty, 1971). Published reports describe hydrogeological problems associated with the Belton Reservoir on the Leon River (Colligan, 1951), Canyon Lake on the Guadalupe River (Webb and Huff, 1962; Clark, 1974; Cook, 1974), and the Amistad Reservoir on the Rio Grande (Kiersch, 1952; Kiersch and Hughes, 1952a,b; Henderson and others, 1959). Numerous other sites along major streams have also been considered for future construction of dams (United States Army Corps of Engineers, 1965).

Speleological Problems

Published research on caves and surficial karst of the Edwards Plateau is varied, but largely concerned with particular phenomena or specific localities. Much of this work is cited and reviewed where appropriate in the discussions of areas selected for detailed study (Chapters 5 through 13). Some additional investigations extend beyond the immediate scope of the present study, but they are relevant to the karst of the Edwards Plateau and are worthy of comment.

Surficial Karst. Surficial karstic landforms on the Edwards are diverse and include solutional and collapsed dolines, karren, lapies, kamenitzas, clints, grikes, caliche, and travertine (Parvin, 1979). Karren lapies, clints, and grikes are common throughout the plateau. Although little is written about them, their significance and origin is similar to whose in other karstic regions (Cvijic, 1024; Bogli, 1960, 1980; Jennings, 1971; Sweeting, 1973; Bleahu, 1974; Jakucs, 1977). However, several other minor solutional forms of the plateau have received more attention. These include (1) beveled clints and pavement, grooves and kamenitzas (Udden, 1925; King, 1927; Smith and others, 1941; Bland, 1958) similar to those reviewed by Lang (1937), Wentworth (1944), and Hedges (1969), and (2) solutionally sculpted and flattened pebbles of limestone (Udden, 1913; Bryan, 1929; Scott, 1947).

Pool (1972a, 1973, 1976, 1977) studied playa lakes of the western Edwards Plateau and discussed their solutional origin with respect to structural control, surficial drainage, and recharge. The playas are similar to those of the Llano Estacado (High Plains) to the northwest (Meigs and others, 19ss; Patton, 1935; Germond, 1940; Reeves, 1966, 1972, 1973, 1974; Reeves and Parry, 1969; Woodruff and others, 1979).

Deposits of caliche are prevalent in the soil and regolith developed on Cretaceous rocks of the Edwards Plateau (Blank and Tynes, 1965; Amsbury, 1968). Some of these are karsted or developed in shallow dolines or playas. Formation of caliche on the plateau is similar to that of well-studied deposits in the neighboring Llano Estacado (High Plains) and Gulf Coast (Brown, 1956; Cooke, 1951; Libby, 1951; McGannon, 1976; Price, 1933, 1940a,b, 1941, 1949, 1958, 1962; Rightmire, 1967; Reeves, 1970, 1971, 1976; Sidwell, 1943; Stenzel and others, 1948) and elsewhere (Goudie, 1973; Goudie and Pye, 1983).

An investigation into the relation of caves and surficial karst to morphometric properties of surrounding landscapes was begun during the course of the present study. Drainage densities of two localities within the highly-dissected southeastern Edwards Plateau have been examined. A review of the technique and initial findings are presented in Appendix F. This method of analysis has since been extended to the karst of the Mississippian Plateau of western Kentucky (Kastning and Kastning, 1980). The study of both areas is continuing at this writing.

Paleokarst. Paleokarst, both buried and exhumed, occurs throughout the Edwards Plateau and within rocks of many ages. Baker (1933), Barnes (1956b), and Barnes and Bell (1954a,b,c) describe deposits of galena within structure of solutional subsidence in the upper Cambrian rocks of the Llano Basin. Cloud and Barnes (1946a), Plummer (1943), and Freman (1962) provide maps and sections of collapsed structures in Ordovician and Carboniferous rocks that contain younger

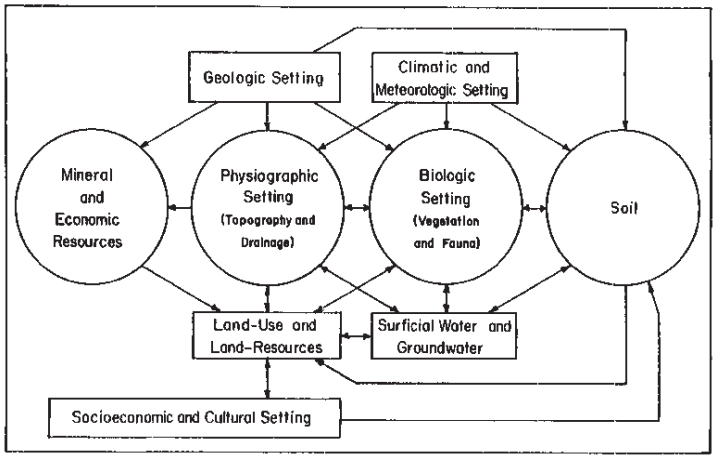


Figure 7: Interrelationships among geologic, biologic, and geographic environments. Arrows portray cause-and-effect relationships.

formations.

Paleokarst also occurs within the cretaceous carbonates of the plateau. Maps of large areas of solutional subsidence in the Devils River Uplift are shown by Freman (1958). Abbott (1975) and Woodruff and Abbott (1979a) discuss relict porosity developed prior to the late Cretaceous within the Edwards Limestone of the San Marcos Arch. Collapse and deformation caused by intrastratal removal of gypsum are described by Rose (1972), Facundus (1968), and Facundus and Moore (1970). Other examples of paleokarst within Cretaceous units are provided by Dumble (1918), Bowser (1927), Adkins (1932), Langford (1942), Livingston and Bennett (1944), George and others (1947, 1952), and Quinlan and Smith (1967, 1968). Some of these findings have been reviewed by Quinlan (1978).

Nitrate Deposits in Caves. Caves of the Edwards have served as sources of nitrate (Phillips, 1901; Gale, 1912, 1932; Mansfield and Boardman, 1932). Although some nitrate was used in the early manufacture of gunpowder, most deposits, in the form of bat-guano, have been mined for organically rich fertilizer. Bracken, Ney, Frio, and Fern Caves harbor tens of millions of bats in the summer and have produced tons of fertilizer during the first half of the century. Mining of guano from caves ceased when potent and inexpensive artificial fertilizers became available and as the largest deposits of guano were depleted.

Paleontological Deposits. Many caves of the plateau are rich with fossil remains of vertebrates (Frank, 1961a,b, 1962a, 1964, 1965b; Lundelius, 1967; Lundelius and Slaughter, 1971). Notable finds include those described by Dalquest and others (1969), Evans (1961), Frank and Frank (1963), Graham (1976), Hill (1966), Homan (1969), Kennerly (1956), Kurten (1963), Lundelius (1960, 1975a,b), Pettus (1956), Semken (1960a,b, 1961, 1967), Slaughter (1964, 1965a,b), Tamsitt (1957), and Weigel (1967).

SUMMARY

The Edwards Plateau is a distinct geographic unit. It is delineated as a physiographic division and as a geologic, hydrologic, climatic, biologic, and perceptual region. The geographic factors discussed in this chapter are strongly interrelated through cause and effect (Figure 7). The geologic and climatic settings are independent variables that, acting together, influence the physical, biological, and cultural settings. The karst of the Edwards Plateau, which evolved within this regional framework, in turn now exerts an influence on the development and use of the land.

2

GEOLOGIC SETTING OF THE EDWARDS PLATEAU

INTRODUCTION

The Edwards Plateau is a distinct geographic region defined by its physiographic, climatic, pedogenic, vegetational and socioeconomic characteristics. The geologic setting has greatly influenced most of these characteristics (Figure 7). This chapter introduces the geology of the Edwards Plateau and presents (1) a summary of the stratigraphy and lithology of the bedrock, (2) a discussion of structural geology and tectonism, and (3) a review of the geologic history. The hydrogeology and geomorphology of the region are treated in subsequent chapters.

STRATIGRAPHY

The area contains a remarkable diversity of bedrock, including sedimentary rocks (limestone, dolostone, marl, chalk, sandstone, siltstone, mudstone, clay, shale, conglomerate, and alluvium), metamorphic rocks (schist, gneiss, quartzite, and marble), and igneous rocks (granite, basalt). All geologic periods except the Triassic and Jurassic are represented by rocks at the outcrop. However, Silurian, Devonian, and Mississippian units are thin and patchy (Plummer, 1943; Barnes and others, 1945, 1966; Barnes, Cloud, and Warren, 1947; Cloud and others, 1957), and Permian rocks are present only as localized exposures in Concho, McCulloch, Crockett, and Menard Counties (Baker and others, 1965; Barnes, 1976a). Tertiary rocks are found only along the southern margin of the area.

A simplified geologic map of the area is shown in Plate 3. Sheets of the Geologic Atlas of Texas at 1:250,000 scale (Barnes, 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b) provide considerably more detail and incorporate current stratigraphic nomenclature. The atlas sheets serve as the geologic base for this study.

Approximately 80 percent of the area is underlain by Cretaceous rocks (Plate 3), and Paleozoic and Precambrian rocks of the Llano Basin account for most of the remaining bedrock. Quaternary sediments occupy many large fluvial valleys and occur locally along low-lying areas immediately south and east of the Balcones Escarpment.

A compilation of most of the known lithic units within the area is presented in Appendix B and includes formal groups, formations, and members as well as several informally named units. The table itemizes type of rock, generalized lithic description, thickness, and extent of outcrop. Units significant to this study are so noted, and principal references are provided for further information.

Precambrian and Paleozoic rocks crop out in the Llano Basin and along the northern boundary of the area. Stratigraphic relations of these rocks are shown in the geologic column of Plate 4. A generalized geologic section of Cambrian and Lower Ordovician rocks (Figure 8) shows these units to be generally continuous across the Llano area. Outcrops of late Ordovician through Mississippian rocks are less extensive, and Pennsylvanian and Permian rocks have their southernmost exposures along the northern margin of the Llano Basin.

Mesozoic rocks are represented by both Lower and Upper Cretaceous sedimentary units. Several paleogeographic and structural elements influenced deposition during the Early Cretaceous (Figure 9). This resulted in a diversity of depositional environments, and, as a consequence, Cretaceous lithic units vary markedly in lithology, thickness, and distribution over the area. Correlation of Comanchean and Gulfian rocks has been accomplished through detailed biostratigraphical analysis. I have compiled a generalized correlation chart for the Cretaceous rocks of the Edwards Plateau based on numerous published stratigraphic studies (Plate 5). The chart clearly shows the proliferation of formal and informal lithic units. It provides correlation among rocks from different regional elements and indicates transitional or unconformable boundaries between successive units. Use of the chart in conjunction with Appendix B and the maps of Plate 3 and Figure 9 will help the reader to identify the geographic and stratigraphic position of Cretaceous rocks discussed in this study.

Cenozoic deposits within the area are limited to one Tertiary clastic formation (Indio Sandstone) along the southern boundary and to Pleistocene and Holocene deposits, such as alluvium, colluvium, windblown sand, deposits associated with terraces, caliche, playas, alluvial fans, marl, and gravel. Most gravel occurs within stream valleys or in deposits beyond the Balcones Escarpment as relict sediment from Tertiary and Quaternary erosion of the Edwards Plateau.

Development of karst in central Texas has been most intense on calcitic and dolomitic rocks of the Lower Cretaceous Series, particularly on the Cow Creek Limestone, the Glen Rose Formation, and the Edwards Group. However, some rocks of the Upper Cretaceous and Paleozoic have also been karstified. Most noteable among these are the Cap Mountain Limestone (Cambrian), limestone and dolostone of the Ellenburger Group (Ordovician), Marble Falls limestone (Pennsylvanian), and rarely, the Anacacho facies of the Austin Chalk (Upper Cretaceous). Some less soluble rocks have influenced cavern development by acting as resistant and protective caprocks or as aquicludes on which flow through conduits has been perched. Minor karst and pseudokarst have developed on rocks of low solubility, including granite. Examples illustrating stratigraphic and lithologic influence on karstification are discussed in Chapters 5 through 12.

STRUCTURAL FEATURES

The structural framework of the Edwards Plateau is shown in Figure 10. Major structural elements include the Llano Uplift, Ouachita Structural Belt, Devils River Uplift, Concho Arch, Edwards Arch, Kerr Basin, Val Verde Basin, Bend Arch, San Marcos Arch, and the Balcones Fault Zone. Most of these have affected development of karst either directly or indirectly (Chapters 5 through 12).

Karstic Paleozoic rocks of the Llano Basin exhibit dips in accordance with the domal structure of the Llano Uplift. Speleogenesis in this region has been locally controlled by faults and joints (Chapters 8 and 9). Exhumed paleokarstic features in the Precambrian and Paleozoic rocks are relicts from successive periods of uplift, erosion, and renewed deposition during the Paleozoic.

Structural elements, as shown in Figure 10, have influenced deposition of terrigenous clastics, carbonates, and evaporites (compare with Figure 9). This has controlled the lithology and stratigraphy of the bedrock, which, in turn, has affected development of karst. Uplift along the San Marcos Arch promoted the development of karst during the Early Cretaceous (Rose, 1968, 1972; Woodruff and Abbott, 1979a; see also Chapter 10). Fractures associated with this arch are detectable on the surface (Leonard, 1977b), as are fractures of the Devils River Uplift, a zone of Late Paleozoic faulting (Leonard, 1977a, 1978; see Chapter 5).The Balcones Fault Zone consists of a series of *en echelon* normal faults, most of which are displaced downward toward the south and southeast (Plate 3 and Figure 10). The fault zone can be traced from northeast of Dallas, past Waco to Georgetown, southwest through Austin, to San Antonio and then west through Uvalde to north of Del Rio. The width of the zone varies from 30 km at Georgetown to 17 km at Austin, 34 km at San Antonio, 27 km at Uvalde, 10 km at Brackettville, and 5 km north of Del Rio. The length and density of individual faults are greatest between Austin and San Antonio. Faults become shorter and fewer west of San Antonio and north of Austin. Maximum, total displacement across the fault zone is approximately 520 m in Hays County (DeCook, 1960, 1963) and decreases westward to 215 m in Uvalde County (Welder and Reeves, 1962, 1964). Maximum displacements along any single fault in these same areas are 220 m and 60 m respectively.

The Balcones Fault Zone has exerted greater control on the geomorphic evolution of the Edwards Plateau and its karst than has any other structural element. Uplift of the Edwards Plateau and displacement along the fault zone provided a new regional baselevel to the southeast, toward which drainage and erosion progressed during the Late Tertiary and the Quaternary. Subsequent incision of streams allowed development of caves. Fractures created by Balcones faulting, as well as individual faults, have strongly modified regional and local hydrogeologic settings and have guided speleogenesis.

Many smaller, structural features have locally affected karstification and

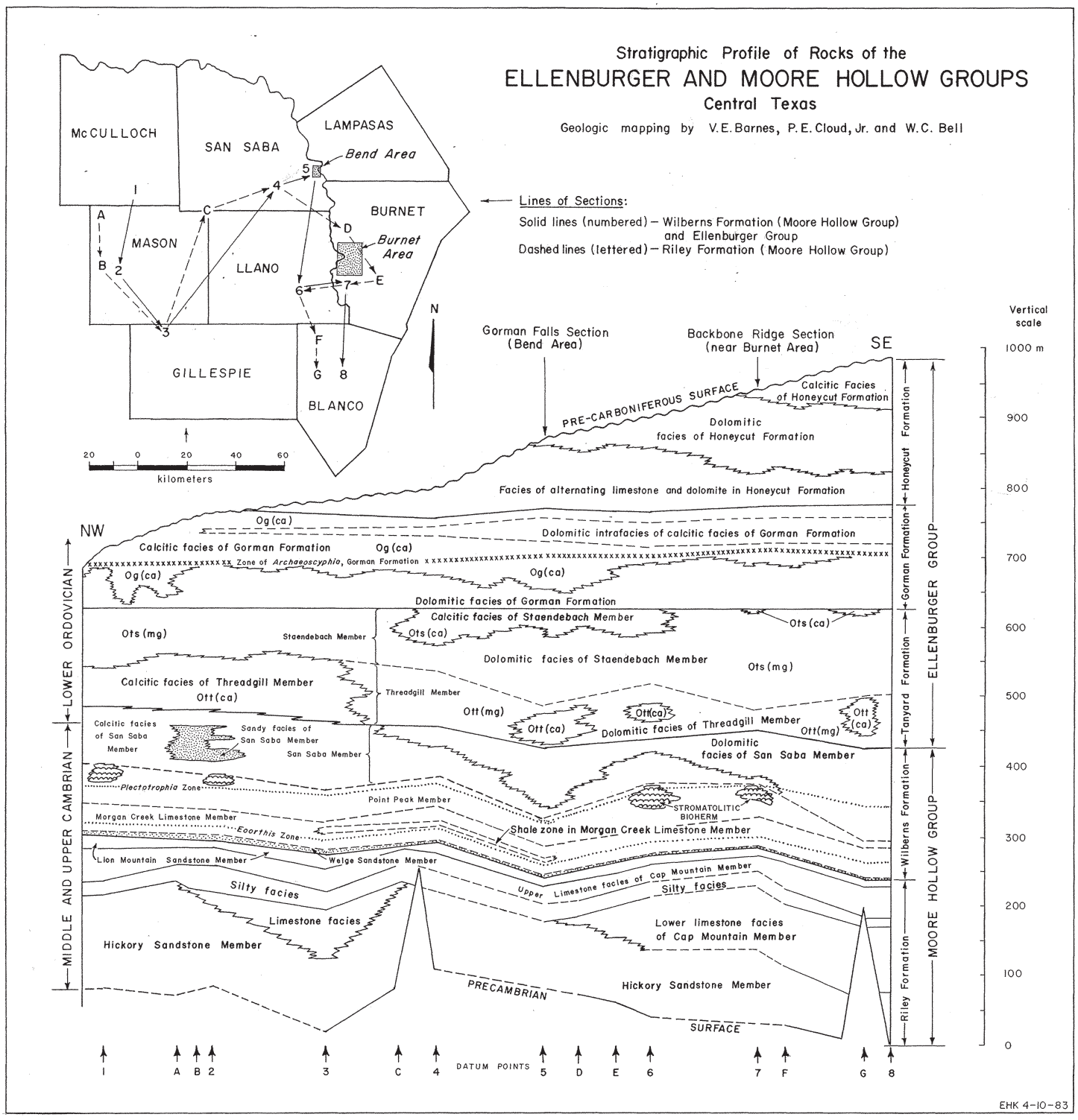


Figure 8: Generalized geologic section of Cambrian and Lower Ordovician rocks of the Llano Basin. Modified from Barnes and Bell (1977, figure 1, p. 8).

speleogenesis. Examples are presented in Chapters 5 through 13. The regional role of structure on the Edwards Plateau karst is summarized in Chapter 14.

SUMMARY OF THE GEOLOGIC HISTORY OF THE EDWARDS PLATEAU REGION

This study is primarily concerned with the development of karst since the Miocene Epoch. However, there are several significant and interesting examples of karst that formed during the Paleozoic and Mesozoic Eras. The geologic history of the area of study given here is a brief summary of depositional, tectonic, and erosional episodes from the Precambrian to the present. It serves as a chronologic framework for discussions that follow in later chapters.

Precambrian Era

Precambrian rocks of central Texas record a single major cycle of sedimentation, deformation, metamorphism, and granitic intrusion (Clabaugh and McGehee, 1972). Thick sedimentary sequences were deposited in a geosyncline extending northwest and southeast through the region. Early deposits included rhyolithic volcanic rocks, tuffaceous sediments, and lesser amounts of feldspathic sandstone, siltstone, and limestone. Magnesian and iron-rich calcareous mud; shallow-water limestone and dolostone; carbonaceous, sulfide-rich mud; clean, white clay and siltstone; and arkosic sandstone were later deposited over the earlier sediments. As the Precambrian seas gradually deepened, terrigenous sediments were laid down, and limestone and mud deposition ceased. The geosyncline underwent deformation, and sediments were folded into elongate anticlines and synclines. Thrust faulting and regional metamorphism at moderate depth ensued. The Valley Spring Gneiss and Packsaddle Schist were the principal metamorphic rocks formed at this time. Mafic and igneous rocks were later intruded, and the Red Mountain Gneiss (originally a granite) was emplaced as a swarm of sills (Clabaugh and Boyer, 1961). As regional metamorphism ended, large granitic intrusions rose into the folded country-rock, further folding and faulting the metasediments. The granitic rocks included, in order of emplacement, the Town Mountain Granite, Oatman Creek Granite, Sixmile Granite, and a quartz porphyry known as Llanite. During the final stages of geosynclinal deformation, metasomatic and retrograde changes occurred locally in metamorphic wall-rocks. Smaller intrusions of granitic magma and injection of small, mafic dikes were the last emplacements of the Precambrian. The next 400 million years (Late Precambrian and Early Cambrian) were characterized by uplift and erosion which destroyed the Precambrian mountains. Less resistant rocks, such as the Packsaddle Schist, some of the Valley Spring Gneiss, and most of the granites were eroded to a generally uniform level, and more resistant rocks, such as the Oatman Creek Granite and the remainder of the Valley Spring Gneiss, formed isolated hills and ridges (Barnes and Bell, 1977). The distribution of Precambrian rocks is shown by Flawn (1956, plate 1).

Paleozoic Era

Paleozoic sedimentation began in Middle Cambrian time as the sea encroached from the southeast. Climate in land areas was semiarid, with little or no vegetation. Evidence for this includes (1) the presence of ventifacts in the Hickory Sandstone (Barnes and Parkinson, 1940), (2) absence or scarcity of the clay fraction in sediments of the Moore Hollow Group, probably owing to wind, (3) chatter marks on grains of sand indicating eolian transport, and (4) relatively little solution of ridges of marble of Precambrian rocks, indicating a general lack of moisture (Barnes and Bell, 1977).

Subsidence continued at a rate equivalent to that of sedimentation during the Middle Cambrian up until the start of deposition of the Cap Mountain Limestone. Increased subsidence and inundation then allowed the silty zone of the Cap Mountain Limestone to be deposited. Subsidence slowed, and the upper, limestone facies of the Cap Mountain was deposited. Subsidence may have ceased entirely upon deposition of the Lion Mountain Sandstone. Erosion of the upper beds of the Lion Mountain occurred as this unit emerged for a short time during the Late Cambrian. Deposits of the Welge Sandstone spread northwestwardly over the eroded Lion Mountain surface and the Morgan Creek Limestone was deposited as a thick sequence above this. Subsidence may have ceased once more. Conditions favorable to stromatolitic growth persisted during the remainder of the Cambrian and Ordovician. Following Morgan Creek deposition the type of sedimentation varies geographically. The basal Point Peak beds are shallow-water deposits in the western area. Subsidence and sedimentation continued concomitantly. Climates were drier and windborn silt is common in the Point Peak. Subsidence exceeded deposition near the end of Point Peak deposition, particularly in the western area. The San Saba Member was subsequently deposited, generally in a shelfal environment. Its facies vary from calcitic to dolomitic to quartzose (Figure 8). The overlying Threadgill Member of the Tanyard Formation was deposited in an intertidal environment. To the east this is less so, and the Threadgill is nearly identical to the San Saba. San Saba-Threadgill deposition marked the close of the Cambrian Period (Bridge and others, 1947; Bell and Barnes, 1961, 1972; Barnes and Bell, 1977).

Lower Ordovician rocks are represented by the Ellenburger Group which includes the Tanyard, Gorman, and Honeycut Formations. The Llano Basin remained relatively stable during the Early Ordovician Epoch. A shallow-marine (intertidal and supertidal) environment persisted, grading from shoal-water conditions at the south and east to slightly deeper waters at the north and east. Changes in depth of water were minor. Penecontemporaneous dolomitization resulted in extensive calcitic and dolomitic facies which can be mapped regionally (Figure 8). During the Early Ordovician the region was environmentally similar to the Bahama Banks off Florida, and soft, pure carbonate muds and warm, intermittently turbid waters prevailed. As a result, units of the Ellenburger limestone are dominantly aphanitic and pelletal, and units of dolostone are microgranular to coarse grained (Cloud and others, 1945; Cloud and Barnes, 1946a, 1957; Barnes and Cloud, 1972).

The Llano region was presumably above sea level during the Middle Ordovician Epoch and perhaps into the Late Ordovician. The only post-Ellenburger Ordovician rock recorded from central Texas is the Burnam Limestone (Upper Ordovician), which is preserved in a karstic, collapsed structure in southern Burnet County. The depositional environment of this unit appears to have been near a wave-breaking reef or near pre-existing coral reefs, indicating generally shallow water in the area (Barnes and others, 1953; Barnes and Cloud, 1972).

Similar geologic conditions existed during the Silurian Period in the Llano region. Two small deposits of the Starcke Limestone have been found in collapsed structures in southern Burnet County. Apparently Upper Ordovician and Silurian deposits were once extensive but have been removed by erosion during any of several episodes of emergence during the Paleozoic Era (Barnes and others, 1966; Barnes and Cloud, 1972).

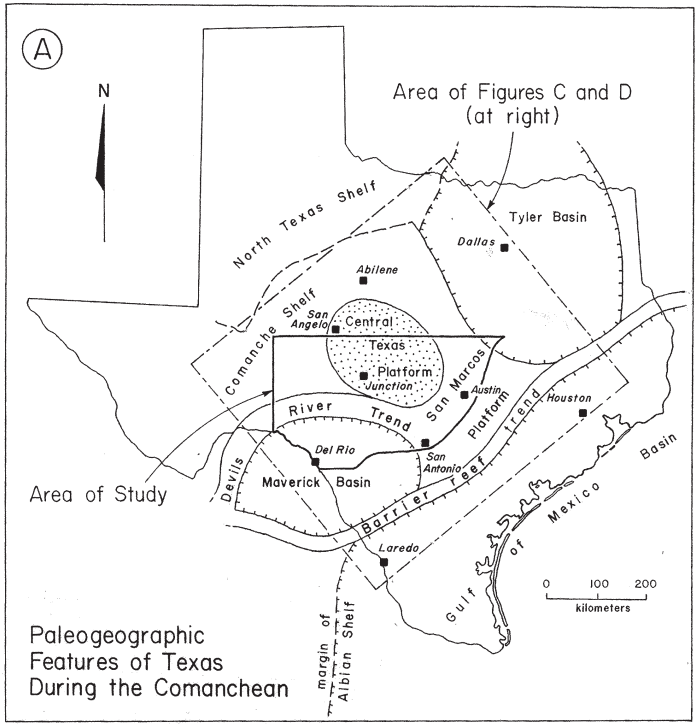
Episodic transgression and regression continued during the Devonian and Early Carboniferous Periods. Six formations have been mapped, but outcrops are small and depositional environments have been difficult to ascertain. Exposures of the Pillar Bluff Limestone, the Bear Spring Formation, and an unnamed limestone occur in collapsed structures or in filled crevasses of karstic origin. Overlap by Devonian marine conditions progressed from east to west across the region. The presence of Devonian formations, some of which were apparently extensive, proves that the Ellenburger rocks of the Llano region were truncated by erosion prior to the Lower Devonian. Truncation probably followed the Middle Ordovician, when at least 200 m of material were removed from the top of the Ellenburger (Barnes and others, 1945; Cloud and Barnes, 1946a; Barnes, Cloud, and Warren, 1947; Cloud and others, 1957; Wilson and Majewske, 1960; Barnes and Cloud, 1972).

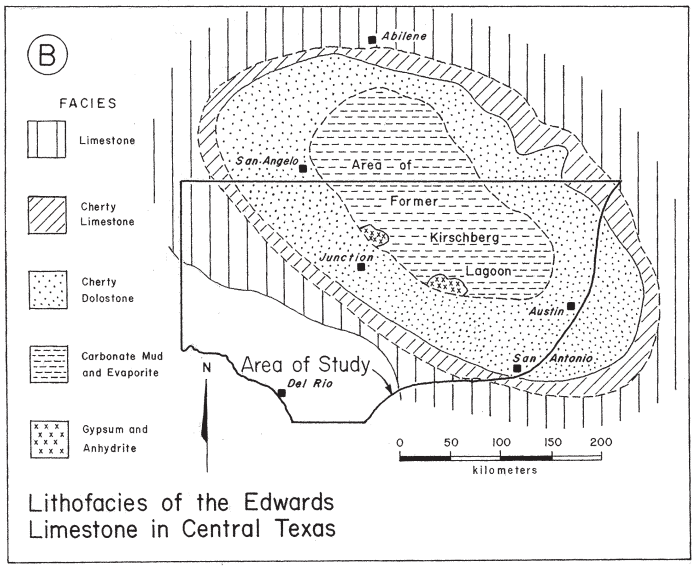
Cloud and Barnes (1946a, p. 109-111) have identified eight significant erosional unconformities among Lower Ordovician, Devonian, and Carboniferous rocks. Most unconformities represent periods of emergence of the Llano uplift, followed by subaerial erosion and marine transgression and sedimentation. Unconformities among Devonian and Lower Carboniferous rocks are difficult to recognize because the units are laterally variable in lithology and have been preserved only in isolated karstic depressions developed on the Ellenburger carbonates during times of emergence.

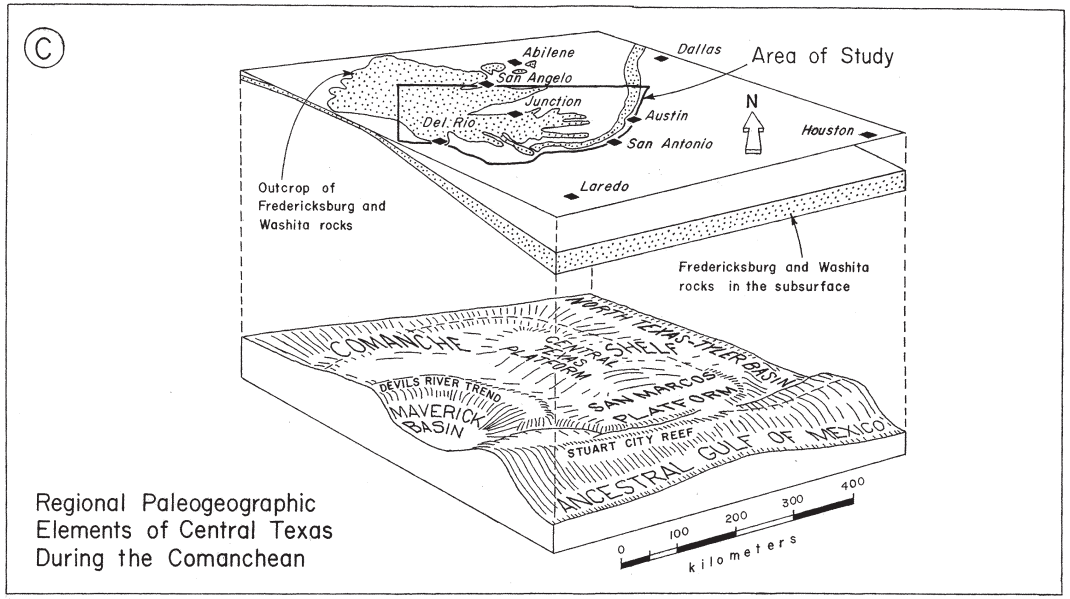
The Houy Formation represents the transition between the Upper Devonian and Lower Mississippian rocks in central Texas. Mississippian rocks are locally discontinuous but are widespread in the Llano region. They are more persistent laterally than the Devonian rocks. Carboniferous strata are generally exposed in three separate areas in the Llano region: (1) the Marble Falls area in southwestern Burnet County, (2) the Mason area in southwestern Mason County and northeastern Kimble County, and (3) an arcuate strip along the northwestern, northern, and northeastern margin of the Llano region. Marine invasions occurred at several times during the Mississippian and Pennsylvanian.

Exposures of the Chappel Limestone are discontinuous and isolated, and many of these are found within structural dolines developed in the Ellenburger rocks during the Mississippian. The other Mississippian unit, the Barnett Formation, is more extensive in outcrop and is unconformable at its base and top. The Barnett strata were truncated and locally completely removed in the late Mississippian (Cloud and Barnes, 1946a; Plummer, 1943; Bell, 1972). The deposition of Ordovician, Silurian, Devonian, and Mississippian rocks within dolines suggests multiple episodes of karstification of the Ellenburger.

Marine invasions in the Llano region continued into the Pennsylvanian Period and the early part of the Permian Period. The Marble Falls Limestone, Smithwick







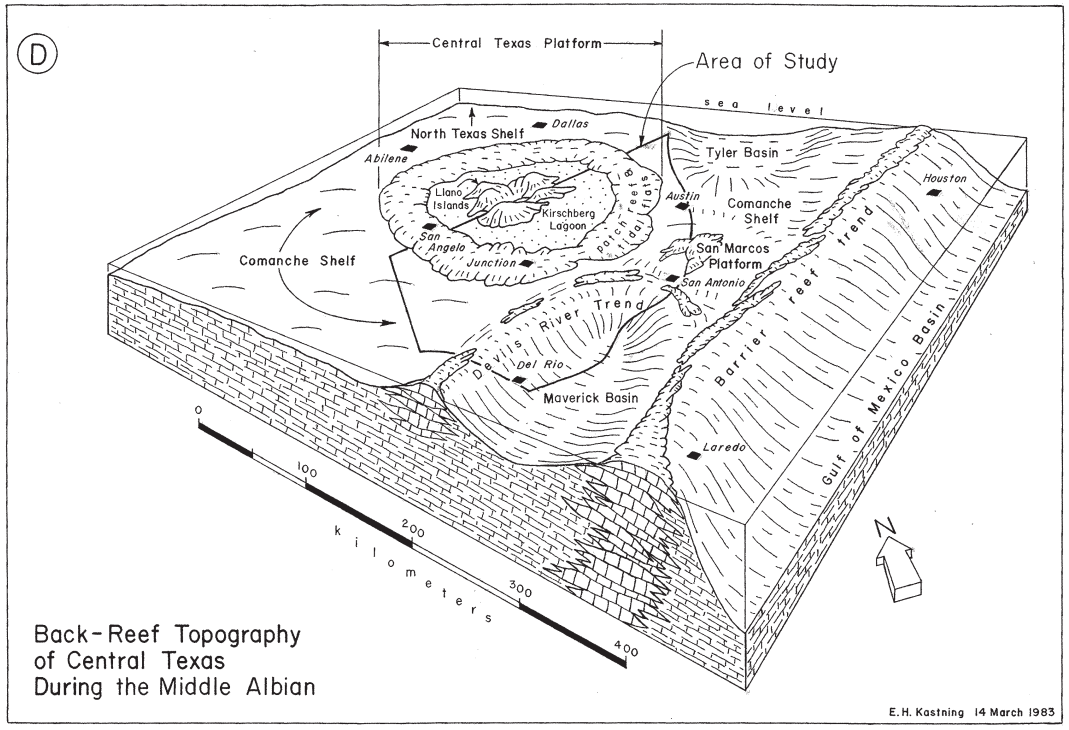


Figure 9: Paleogeographic and structural elements of central Texas during the Early Cretaceous and composite map of litholaies. Modified from Fisher and Rodda (1969, figure 14, p. 70), Rose (1972, figure 2, p. 5), and Young (1972a, figures 2 and 3, p. 4–5).

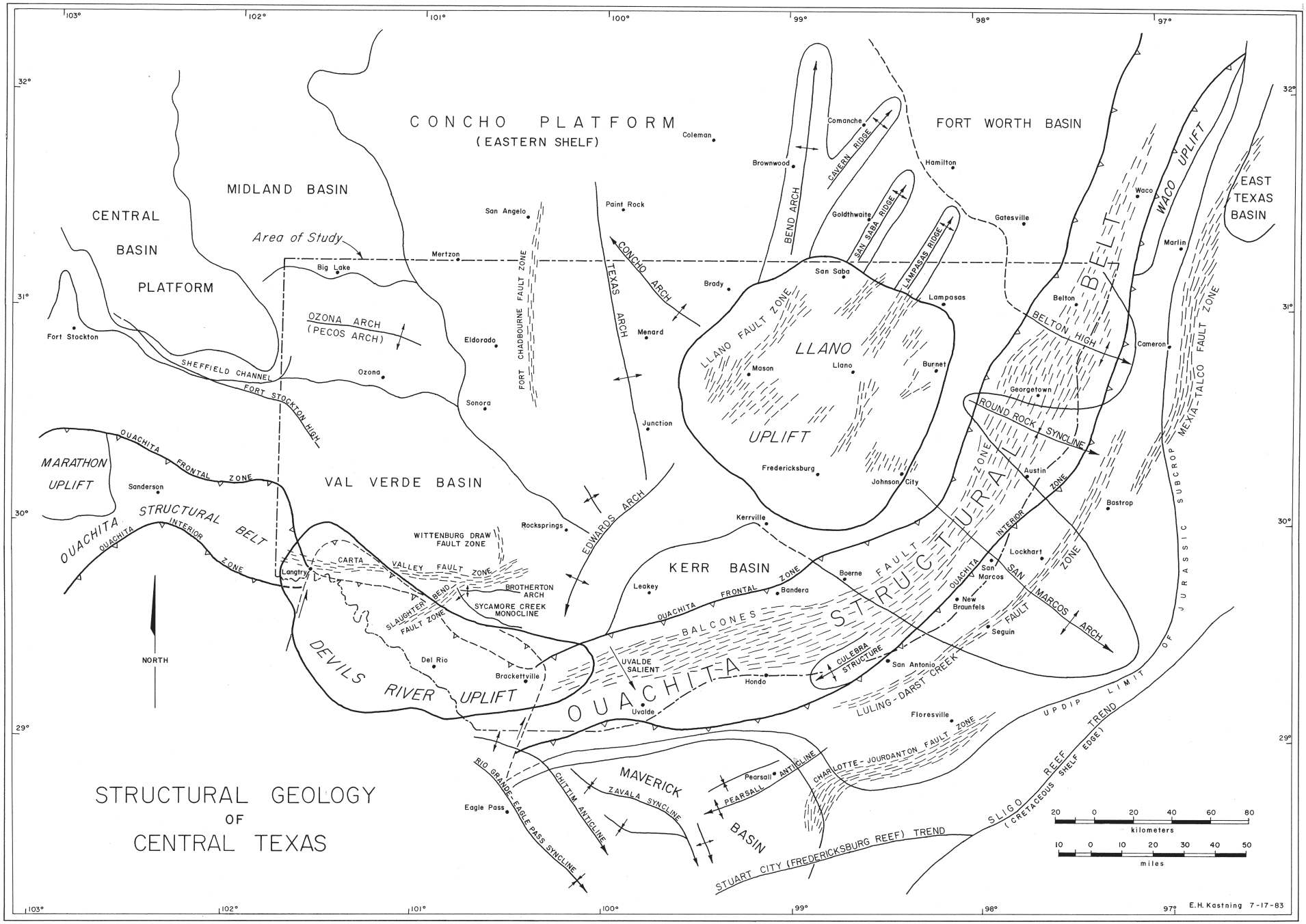


Figure 10: Composite map of major structural and tectonic elements of central texas. Sources listed in Appendix B.

Shale, and numerous formations and members of the Strawn, Canyon, and Cisco Groups were deposited during this time. Unconformities resulting from emergence and truncation separate many of these units. Undisturbed beds of the Canyon Group overlap faulted strata of the Tanyard Formation. Faults in the Tanyard also cut the Marble Falls Limestone and Smithwick Shale elsewhere. This suggests that major faulting occurred in the Llano region between Smithwick and Canyon deposition during the Pennsylvanian (Plummer and Moore, 1921; Plummer, 1943; Cloud and Barnes, 1946a).

Epeirogeny during the Pennsylvanian Period brought the Llano region well above sealevel. This was accompanied by the orogenic phase of the Ouachita System along the eastern margin of the area of study (Powers, 1928; Van der Gracht, 1931a,b; Cheney and Goss, 1952; Flawn, 1956, 1959a; King,1961). Keller and Cebull (1973) interpreted the Ouachita System as an extension of the Appalachian orogen and applied a cordilleran-type model to explain its origin in light of global tectonics.

Mesozoic Era

Central Texas underwent extensive erosion during the Permian, Triassic, and Jurassic Periods, exposing Precambrian and Paleozoic rocks. The resulting erosional surface, termed the Wichita paleoplain by Hill (1901), was later covered by a transgressive Cretaceous sea. The ancestral basin of the Gulf of Mexico sank, and its northern flank tilted southward during Cretaceous deposition (Young, 1972b). Each Cretaceous unit up to the Edwards Group (Plate 5) is overlapped by younger formations, each of which in turn rests on exposed Precambrian and Paleozoic rocks. The first units deposited were the Sycamore and Hosston Sands. Local uplift resulted in an erosional surface. Fluvial and marine environments followed and the Hammett Shale and Cow Creek Limestones overlapped the Sycamore. Sinking along the Gulf Coast accompanied further transgression, and the Hensell Sand and Glen Rose Formation were deposited. The sea had extended part way around the Llano region at the close of Glen Rose deposition. Sedimentation was controlled by the Stuart City reefal complex to the southeast during Trinity deposition and up until the end of Washita deposition (Figure 9). The Glen Rose represents deposition dominated by fine-grained dolomitic limestone, fine-grained limestone, and evaporites in shallow water (Young, 1972b). The contact between the Trinity and overlying Fredericksburg rocks near the Llano region may be locally disconformable (Moore, 1961a,b). In the northeastern part of the area the Paluxy Sandstone overlying the Trinity rocks interfingers with the lowermost Walnut beds (Moore, 1961b).

Marine transgression proceeded during Fredericksburg deposition, keeping pace with subsidence along the Gulf Coast. Some of the Llano region became isolated as islands by at least middle Comanche Peak deposition, as the sea crossed the Bend Arch and Callahan divide from the east and advanced northward around the western side of the Llano region. However, Glen Rose deposition occurred locally in pre-Cretaceous valleys north of Brady (McCulloch County) and west of the Bend Arch. The entire area of study was inundated by the end of Fredericksburg deposition. Regional dips and variations in thickness of formations indicate that the Llano area was covered by Fredericksburg and Washita rocks (Young, 1972b).

The San Marcos Platform subsided at a lesser rate than did other areas of the Comanche Shelf during Fredericksburg and Washita deposition (Rose, 1968, 1972). Only about 150 m of shallow-marine and tidal-flat deposits accumulated on the San Marcos platform, while nearly 300 m of grainstone and rudistid boundstone were formed on the Devils River Trend. The San Marcos Platform was subsequently uplifted, and more than 30 m of the uppermost beds of the Edwards Group were removed by subaerial erosion during part of the Washita interval (Rose, 1968, 1972; Woodruff and Abbott, 1979a).

Subsidence continued in the geosyncline, and Georgetown beds were deposited in the northeast part of the study area and on the San Marcos Platform. The Del Rio Clay was laid down as a blanket deposit off the Llano Uplift. Deposition of the Buda Limestone marked the close of the Comanchean. The Buda was presumably not deposited over the Llano region (Hixon, 1959), but it is not known whether the Llano region was under water during this time (Young, 1972b).

Sedimentation ceased during Woodbine deposition (Late Cretaceous) except in the northeast where the Pepper Shale was deposited. The Llano region was forested during Eagle Ford time as indicated by fossil logs in the Bouldin Flag Member of the Lake Waco Formation. The Llano Uplift and San Marcos Platform prevented transport of Eagle Ford terrigenous clastics (derived from the Ouachita Mountains) to the south and west. This allowed the Boquillas Flags to be deposited here instead. Fossil faunas northeast of the Llano Uplift and San Marcos Platform are typically northern European. Faunas southwest of the Diablo Platform are typically Tethyan. Between these two elements the faunas are mixed (Young, 1972b).

Terrigenous sedimentation ended with Eagle Ford sedimentation. The Comanche Platform was maintained in an environment of higher energy during Austin and Taylor deposition. That these deposits once covered the entire Edwards Plateau is suggested by the abundance of Upper Cretaceous fossils in the Tertiary deposits of the Gulf Coastal Plain. The volume of reworked and transported Upper Cretaceous material is so great that exposures of Austin and Taylor units on the Edwards Plateau, north and west of the Balcones Escarpment, must have been extensive at the onset of erosion in the Miocene. Deposits of Austin and Taylor units are now generally absent north and west of the escarpment, except for a few outliers in small grabens along the eastern edge of the Edwards Plateau (Young, 1972b).

Volcanism occurred during the Late Washita to Navarro stages (Early Cenomanian to Maestrichtian) along a belt parallel to and just south and east of the Balcones Escarpment (Dawson and others, 1927; Moody, 1949; Spencer, 1969; Baldwin and Adams, 1970). Volcanic activity was most prevalent during Austin deposition and predated Balcones faulting by about 45 million years.

Cretaceous deposition ended with the formations of the Navarro Group. These beds may also have been continuous over the area of study, but they are now absent on the Edwards Plateau.

Cenozoic Era

Deposition continued during the Tertiary Period. However, the extent of these deposits over the Edwards Plateau can not be fully ascertained. Paleocene marine, near-shore deposits as well as some terrigenous sediments may have covered some of the Edwards Plateau east of the Llano region. Oligocene and Eocene rocks may have originally extended further updip (to the northwest), but they probably did not cover the Llano region. The source of some sediments during Tertiary deposition was the Llano region; however, because it was low in altitude, it did not contribute large quantities of sediment until uplift occurred during the Miocene (Young, 1972b).

Most, if not all, of the movement on the Balcones Fault Zone occurred during early Miocene (Weeks, 1941, 1945a,b; Young, 1972b). Evidence for this includes an abundance of reworked Cretaceous fossils (foraminiferans and molluscan fragments) and clasts of limestone and marl from the Austin Chalk and Taylor Group in the fluvial sandstone beds of the Miocene Oakville Formation (Applin and others, 1925; Plummer, 1932; Renick, 1936; Frizzell, 1954; Wilson, 1956, 1962; Ely, 1957). The Llano region and San Marcos Platform were raised during faulting, while the coastal plain area subsided. An increase in relief and lowering of regional baselevel resulted in removal of most of the Upper Cretaceous and Paleocene rocks from the raised areas and redeposition of this material in the basin of the Gulf Coast to the southeast (Bailey, 1923; Weeks, 1941, 1945a; Young, 1972b). Readjustment of the gradient during the Pliocene or Pleistocene may be inferred from the siliceous Uvalde Gravel found well above present elevations of streams just south and east of the Balcones Escarpment.

This study is largely concerned with the geomorphic evolution of the karst of the Edwards Plateau since the Miocene. The geologic history during the Late Tertiary and Quaternary is discussed, as it pertains to specific sites on the plateau, in Chapters 5 through 13.

3

HYDROGEOLOGIC SETTING OF THE EDWARDS PLATEAU

INTRODUCTION

An understanding of the hydrogeological setting of the Edwards Plateau is essential to the study of its karst. Data collected by the United States Geological Survey and Texas Department of Water Resources are used here to define the climate, flow of streams, and circulation of groundwater within the region and to analyze the Edwards Plateau as a hydrogeologic system.

HYDROLOGIC CYCLE

The hydrologic cycle may be divided into five components: (1) precipitation, (2) evaporation and transpiration, (3) runoff, (4) infiltration, and (5) storage (Figure 11). Precipitation on the Edwards Plateau occurs predominantly as rain; however, some atmospheric moisture enters the system in the form of mist, dew, frost, sleet, snow, and hail. Regional rates of precipitation are summarized in Chapter 1. Evaporation and transpiration are significant factors in the hydrologic budget, and rates vary markedly over the region (Chapter 1). Surficial runoff on the Edwards Plateau, particularly on lower-order streams, is flashy, in response to sudden heavy storms (Chapter 1).

A significant amount of precipitation and runoff enters the ground as infiltration. Some of this is retained in soils, but because soils of the Edwards Plateau are generally thin, most infiltration passes through to the underlying bedrock. Most groundwater on the Edwards Plateau is contained within the three major and two minor, regional aquifers (Figures 4 and 5).

Surficial water and groundwater may be retained over varying periods of time as storage. Some surficial waters are stored in lakes, marshes, reservoirs, and in lesser impoundments such as stock ponds and tanks. Groundwater may reside for long periods (up to tens of years, based on studies of the Edwards Aquifer by Pearson and others (1975)) within the saturated zones of the larger aquifers.

The hydrogeology of the Edwards Plateau karst is concerned with systems of both surficial water and groundwater. However, the geomorphic history of the karst is intimately related to the development of subsurficial circulation and to the evolution of aquifers. In order to properly evaluate paleohydrologic events, it is first necessary to review the present-day hydrogeology of the aquifers of the plateau.

GROUNDWATER

Major Aquifers

The most important aquifers of the Edwards Plateau are the Edwards-Trinity (Plateau) Aquifer and the Edwards (Balcones Fault Zone) Aquifer (Figure 4). Together, these underlie most of the area of study and account for the greatest part of its groundwater-budget. They are entirely separate aquifers and have distinctly different hydrogeologic settings (Figure 11).

Edwards-Trinity (Plateau) Aquifer

The Edwards-Trinity Aquifer underlies the interior of the Edwards Plateau (Walker, 1979; Rees and Buckner, 1980), and its boundaries are generally coincident with those of the outcrop of rocks of the Edwards Group (Figures 4 and Plate 3). The Edwards-Trinity Aquifer is a system of several unconfined (watertable) aquifers. Recharge resulting from precipitation is generally diffuse and through fractures and solutional openings in uplands. Groundwater moves along fractures and bedding-plane partings toward the south and southeast and discharges at springs and seeps along walls of valleys and along streams at baselevel. Points of discharge are most numerous in areas where the surface has been greatly dissected. Springs and seepage account for the baseflow of streams draining the interior of the plateau.

Groundwater flows principally within rocks of the Edwards Group. However, in areas of deep dissection to the east and south, some water is communicated to the underlying Glen Rose Formation and transmitted to springs within that unit (Walker, 1979).

Water is pumped from the Edwards-Trinity Aquifer through numerous small wells that supply ranching and domestic needs. Quality of the water is generally high (Walker, 1979).

Edwards (Balcones Fault Zone) Aquifer

The Edwards Aquifer consists of two distinct parts: (1) a zone of recharge along the trend of the Balcones Fault Zone and (2) a confined (artesian) zone in the subsurface below and just to the south and east of the fault zone (Figures 11 and 12). The boundaries of the zone of recharge are generally coincident with the outcrop of the Edwards Group along the Balcones Fault Zone (Figure 4 and Plate 3). The Edwards Aquifer extends for approximately 400 km along an arc from Del Rio in Val Verde County to Bell County. The central segment of the aquifer, which lies between two groundwater-divides, one near Brackettville in Kinney County and the other near Kyle in Hays County, has been known as the “Edwards Aquifer” in reference to water-supply for the greater San Antonio area (Garza, 1962). The Texas Department of Water Resources now includes the northeastern segment, from Kyle to Bell County, within the contiguous Edwards Aquifer for the purposes of studying and managing the supply of water.

The zone of freshwater of the Edwards Aquifer varies in width from 8 to 64 km. It is bounded on the north and west by the faulted outcrop of the Edwards Group and on the south and east by the “bad-water line”, an interface between fresh and saline water (Garza, 1963; Maclay and Small, 1976, p. 4). The zone of freshwater of the Edwards Aquifer is one of the most productive aquifers in the southwestern United States, supplying all water for several municipal areas and communities and for many irrigated farms. Previous investigations of the Edwards Aquifer are cited in Chapter 1.

General Hydrology and Geology. Maclay and Small (1976, p. 11-15, 1979, 1983) have reviewed the hydrogeology of the Edwards Aquifer. The following paragraphs provide a brief summary.

The Edwards Aquifer is recharged by infiltration at the outcrop of the Edwards Group (Figure 12). Streamflow from the area of catchment on the Edwards Plateau to the north and west enters the highly permeable, exposed limestone along the channels crossing the outcrop. Some recharge occurs in the interfluvial areas where much of the precipitation descends directly to the watertable. Most surficial inflow to the zone of recharge and all surficial outflow from it is measured at gaging stations. The difference between total inflow and total outflow is the calculated recharge. Where areas are not gaged and for runoff within the zone of recharge, runoff is computed from estimates of unit runoff and distribution of rainfall.

Annual recharge to the Edwards Aquifer ranged from approximately 54 x 10^6 m^3 in 1956 to more than 2100 x 10^6 m^3 in 1958 (Garza, 1962, p. 13). Average annual recharge was 692 x 10^6 m^3 from 1934 through 1983.

Regionally, groundwater moves through the aquifer from west to east (Figure 12). The potentiometric surface within the aquifer is shown in Figure 13. Values of head are higher in the area of recharge in the Balcones Fault Zone, where water enters the aquifer, and they are lower in the zone of natural discharge in the eastern part of the aquifer near San Antonio. Some flow is presently diverted toward areas of high rates of pumping, particularly in Bexar County. Pumping from a zone of relatively low transmissivity in southeastern Uvalde County has locally depressed potentiometric contours.

All major springs in the vicinity of San Antonio occur on major faults (Guyton and Associates, 1979). Some faults, especially those that strike northeastwardly, strongly affect the direction of flow of groundwater.

Most regional underflow in Medina County travels eastwardly within blocks bounded by the Cliff Fault on the northwest and the Castroville Fault on the southeast. Yields from wells southeast of the Castroville Fault are generally less than those of wells northwest of the fault. High concentrations of sulfate and rations of concentrations of dissolved ions in the water of the aquifer in southern Medina County suggest that circulation of groundwater is slow.

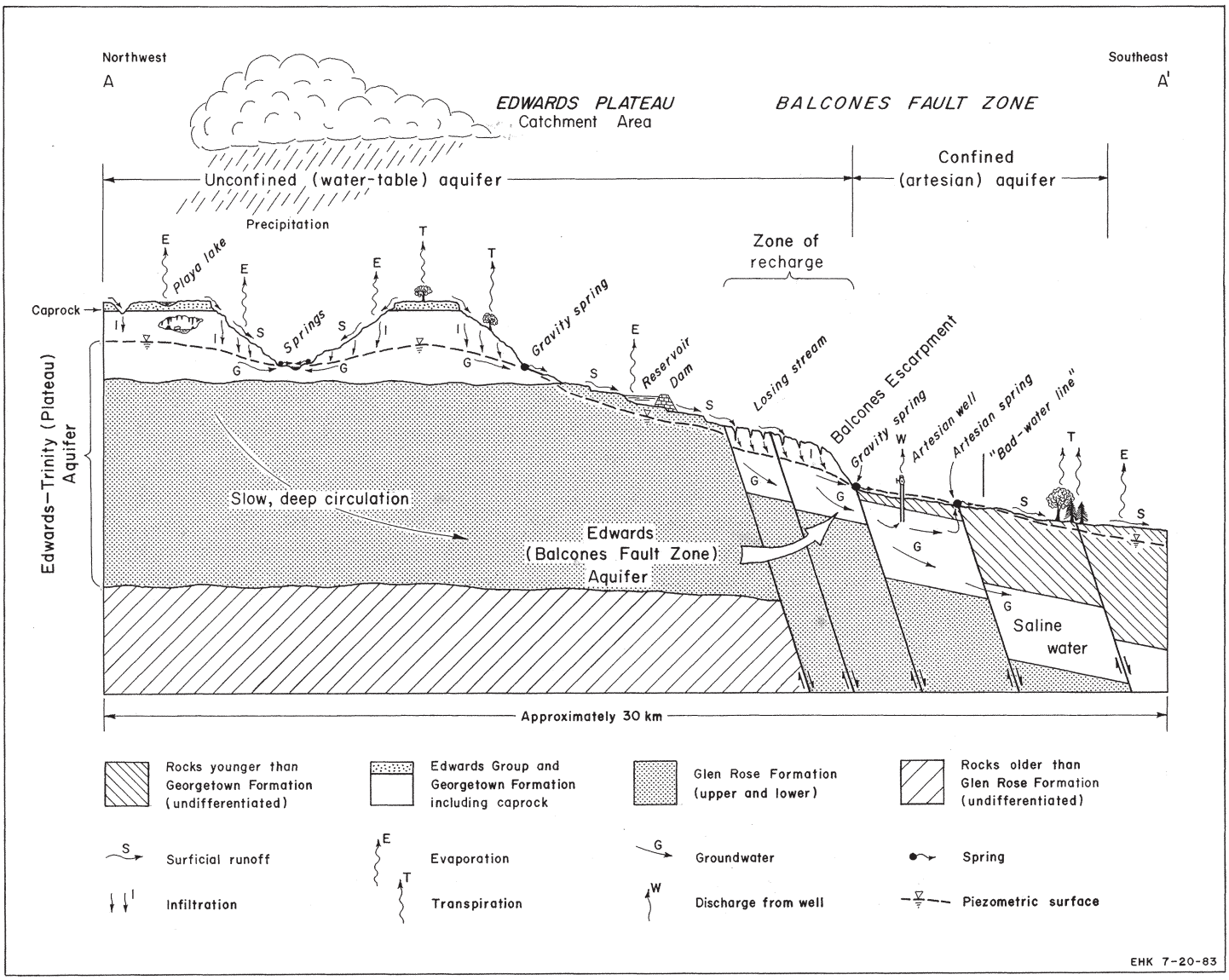


Figure 11: Schematic cross-section of the Edwards Plateau showing components of the hydrologic budget and elements of the Edwards-Trinity (Plateau) and Edwards (Balcones Fault Zone) Aquifers. Typical location section is shown in Figure 12.

Regional underflow in Bexar County travels northeastwardly through a highly faulted zone and toward the fault-blocks bounded by the Comal and Hueco Springs Faults in Comal County. Most flow of groundwater from Bexar County into Comal County is transmitted through the area near the Comal Springs Fault.

The displacement of the Comal Springs Fault is generally sufficient along most of its length to interrupt the continuity of the aquifer across the fault-plane; that is, between water-bearing rocks of the upthrown and downthrown blocks. Similarly, the displacement along the Hueco Springs Fault is sufficient to disrupt the hydrogeological continuity of the upper part of the Edwards Group (Person Formation) from the lower part (Kainer Formation) (George and others, 1947, 1952; Maclay and Small, 1976; Guyton and Associates, 1979).

The Comal Springs and Hueco Springs Faults converge near San Marcos Springs in Hays County (Guyton and Associates, 1979). A transverse fault cutting the northern end of the zone of faults may partially disrupt the rocks of the Edwards Aquifer for approximately 3.2 km northeast of San Marcos Springs. The thickness of the Edwards Aquifer is maximal in the area between the Comal Springs and Hueco Springs Faults, but in some places it is only partially saturated. Northeast of Comal Springs the bad-water line locally crosses the Comal Springs Fault. This might explain a reduction in transmissivity in the aquifer owing to generally lower permeabilities downdip from the bad-water line.

Much of the Edwards Group has been erosionally removed from the faulted area lying between the Hidden Valley and Hueco Springs Faults. In fact, the underlying Glen Rose Formation is exposed in some places north of Bat Cave Fault. Much of the groundwater in the Edwards Aquifer within the area west of the Hueco Springs Fault and southwest of the Guadalupe River moves locally to seepage-springs at or near the contact between the Glen Rose Formation and the Edwards Group. However, some water flows northeastwardly toward Hueco and San Marcos Springs.

In the area north of the Hueco Springs Fault in east-central Comal County the rocks of the Edwards Group are not highly faulted. Here, the flow of groundwater toward San Marcos Springs is relatively unobstructed, and the permeability of the strata largely controls the direction of flow.

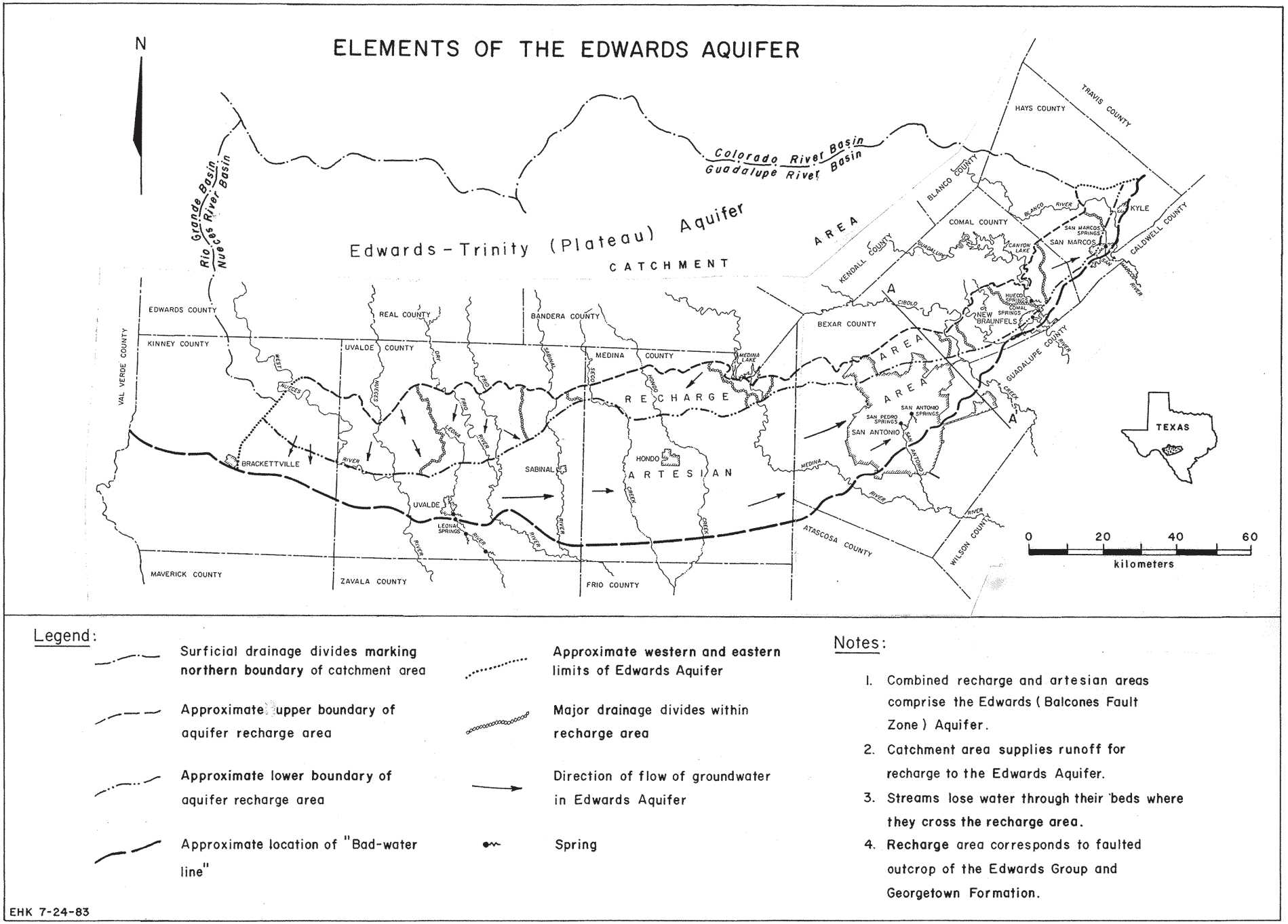
Discharge of groundwater from the Edwards Aquifer is through springs and pumped wells. A relatively small but undetermined amount of discharge may occur as underflow across the bad-water line or as leakage into strata overlying rocks of the Edwards Group. Records of springflow and pumpage maintained by the United States Geological Survey indicate that the average annual discharge from 1934 to 1973 was 683 x 10^6 m^3. The average annual combined discharge of the springs from 1945 through 1974 was 424 x 10^6 m^3, or approximately 62 percent of the average annual discharge. Annual springflow ranged from 86 x 10^6 m^3 in 1956 to about 651 x 10^6 m^3 in 1973. Comal Springs in Comal County (Figure 14), collectively the largest spring in the southwestern United States, ceased to flow for a short time during the summer of 1956. However, San Marcos Springs in Hays County have flowed continuously during the period of record. The flow from Hueco Springs in Comal County often ceases or is much reduced during dry periods (see below for further comments on discharge and hydrographs of these springs.

Pumpage from wells in the Edwards Aquifer is steadily increasing owing to an expanding demand for water for irrigation and municipal needs. Records of the United States Geological Survey indicated that since 1967, discharge from wells

Table 6: Hydrogeologic Characteristics of Subdivisions of Aquifers in Berar County[a]

Aquifer subdivision[b]	Thick-ness	Total porosity[c] (percent)	Matrix permea-bility	Fractures	Description of carbonate facies and pore spaces	Yield to wells
Georgetown Formation	6-12	5	very Low	few, closed	Dense, shaly limestone; mudstone and wackestone; isolated fossil molds.	Yields very little water
Marine and Cyclic Members of Person Formation	24-30	5-15	low	many, open	Hard, dense, recrystal-lized limestone; mud-stone; rudistid biomi-crite; some moldic porosity, locally honey-combed.	Yields moderate to large amounts of water.
Leached and Collapsed Member of Person Formation	18-27	5-20	low to high	many, open	Recrystallized, leached limestone; burrowed mudstone and wackestone, highly leached in places; solution breccias, vuggy, honeycombed, some cav-ernous porosity.	Yields large amounts of water
Leached and Collapsed Member of Person Formation	18-27	5-20	low to high	many, open	Recrystallized, leached limestone; burrowed mudstone and wackestone, highly leached in places; solution breccias, vuggy, honeycombed, some cav-ernous porosity.	Yields large amounts of water
Regional Dense Member of Person Formation	5-7	< 5	very low	closed	Very dense shaly lime-stone, shaley to wispy, dense; mudstone; no open fractures.	Yields very little water
Grainstone Member of¨ Kainer Formation	15-18	5-15	low to moder-ate	few, open	Hard, dense limestone; chalky to hard cemented miliolid grainstone with associated beds of mud-stone and wackestone; locally honeycombed in burrowed beds.	Yields moder-ate amount of water.
Grainstone Member of¨ Kainer Formation	15-18	5-15	low to moder-ate	few, open	Hard, dense limestone; chalky to hard cemented miliolid grainstone with associated beds of mud-stone and wackestone; locally honeycombed in burrowed beds.	Yields moder-ate amount of water.
Dolomitic Member of¨ Kainer Formation	34-46	5-20	low to high	many, open	Hard, dense, dolomitic limestone, recrystal-lized from dolomite, honeycombed in a few burrowed beds; more cavernous in upper part.	Yields moder-ate to large amounts of water.
Kirschberg Evaporite of Kainer Formation	15-21	5-25	low to very high	undeter-mined	Limestone and leached evaporitic rocks with boxwork porosity; solu-tion breccias, most porous and permeable subdivision, locally cavernous.	Yields large amounts of water.

[a]Modified from Maclay and Small (1976, p. 22-25)
[b]Stratigraphic position shown in Plate 5 (San Marcos Platform).
[c]Based on visual examination of cores (Maclay and Small, 1976).



Legend:

- Surficial drainage divides marking northern boundary of catchment area
- Approximate upper boundary of aquifer recharge area
- Approximate lower boundary of aquifer recharge area
- Approximate location of "Bad-water line"
- Approximate western and eastern limits of Edwards Aquifer
- Major drainage divides within recharge area
- Direction of flow of groundwater in Edwards Aquifer
- Spring

Notes:

1. Combined recharge and artesian areas comprise the Edwards (Balcones Fault Zone) Aquifer.
2. Catchment area supplies runoff for recharge to the Edwards Aquifer.
3. Streams lose water through their beds where they cross the recharge area.
4. Recharge area corresponds to faulted outcrop of the Edwards Group and Georgetown Formation.

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Figure 12: Hydrologic setting of the Edwards Aquifer showing areas of catchment, recharge, and artesian conditions. Modified from Maclay and Small (1976, figure 1, p. 5).

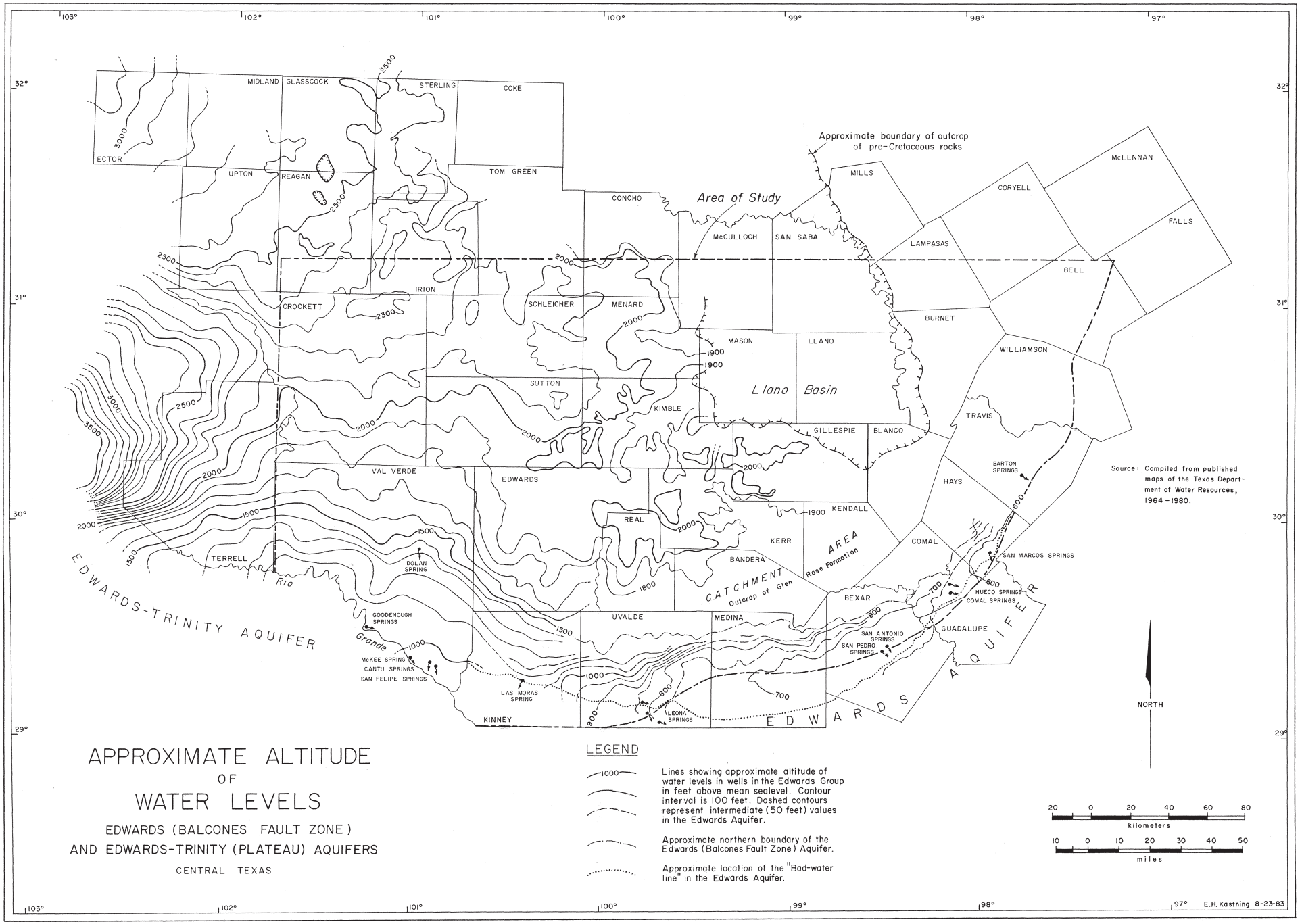


Figure 13: Approximate altitude of water levels in the Edwards and Edwards-Trinity Aquifers. Compiled from Alexander and others, (1964, plate 7), Rees and Bruckner (1980, figure 7, p. 17, and Walker (1979, figure 15, p. 51, 53).

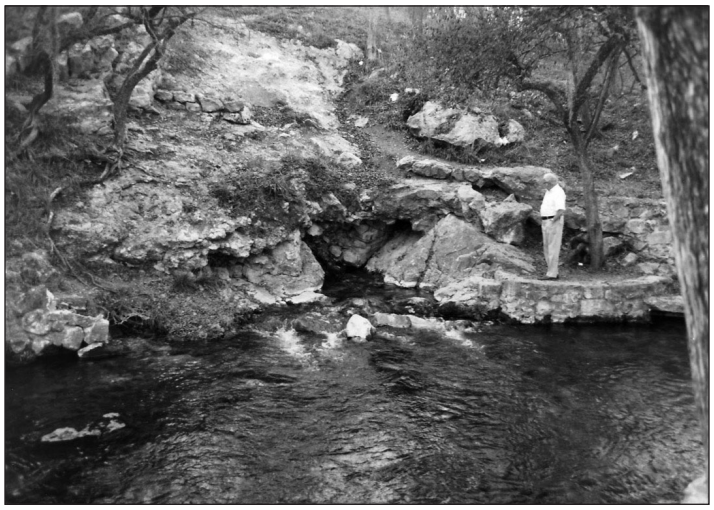


Figure 14: Southwesternmost spring at Comal Springs, New Braunfels, Comal County.

has exceeded 370 x 10^6 m^3 per year except during 1968, which was a wet year. Pumpage in 1971 was 502 x 10^6 m^3 , one of the highest amounts of record. This amounted to approximately 73 percent of the average annual recharge.

The total storage-capacity of the Edwards Aquifer is unknown. Petitt and George (1956) and Garza (1966) estimated the storage-capacity of that part of the aquifer through which waterlevels have fluctuated during the period of historical records. Their estimates range from 15 x 10^6 m^3 to 21 x 10^6 m^3 of decline in head. These estimates are based on the observed relation between changes in waterlevel in an index-well and the difference between annual recharge and annual discharge.

The capacity of the aquifer to transmit water is very high in the region between the outcrop of the Edwards Group and the bad-water line. Evidence for this includes (1) occurrence of many wells of high yield distributed throughout most of the artesian area, (2) rapid and widespread transmission of the effects of pumping, (3) low gradient of the potentiometric surface, and (4) response of waterlevels in some wells of Bexar County to the Alaskan earthquake of 1964 (Montgomery, 1964). Moreover, municipal and agricultural wells in Bexar County commonly yield several thousands of liters per second with resulting drawdowns of only a few meters or less. No precise values of transmissivity are currently available; however, in some areas transmissivity may exceed 124,500 m^2 per day, and at least most places within the artesian zone in Medina, Bexar, Comal, and Hays Counties, the transmissivity probably exceeds 12,450 m^2 per day, Within the unconfined zone and downdip from the bad-water line the transmissivity is considerably less (Maclay and Small, 1976).

Structure of the Edwards Aquifer. The Balcones Fault Zone has interrupted a broad homocline that extends from the nearly flat-lying exposed rocks of the Edwards Plateau to the subsurface of the Gulf Coastal Plain. South and east of the Balcones Fault Zone the regional dip becomes steeper, and rocks of the Edwards Group are buried by more than 3000 m of younger deposits a few tens of kilometers southeast of the bad-water line,

Faulting increases in intensity and complexity eastward from Kinney County to San Antonio. Faults in Bexar County are generally of very high angle but are locally tilted. Because of *en echelon* and stepped faulting, differential movement has placed beds of varying porosities and permeabilities in juxtaposition. This has locally disrupted hydraulic continuity and compartmentalized the Edwards Aquifer (Abbott, 1977a; Maclay and Small, 1979, 1983). Transverse, reverse, and antithetic faults are also found as local components of the Balcones Fault Zone (Maclay and Small, 1976, p. 21).

The Edwards Aquifer is highly fractured. Extensive joints associated with Balcones faulting are oriented parallel to the trend of faults. There are also numerous joints of lesser extent and frequency that strike orthogonal to this trend (Wermund and others, 1978; Kastning, 1980e, 1981b, 1984). In the confined (artesian) zone of the aquifer, movement of water is along the strike of the fault-zone and parallel to regional equipotential lines rather than perpendicular to them. This is particularly true in Medina and Bexar Counties (Figure 13) and had been observed in Comal County by Bills (1957). Arnow (1959) explained this apparent anomaly by suggesting that conduits developed along open fractures that parallel the trend of faulting, and significantly increased the transmissivity of the limestone in this direction. Abbott (1975), on the other hand, regarded the anomaly as a problem of scale. He suggests that more observation-wells spaced closer together would show that the correct hydraulic gradient is actually parallel to the trend of faults. Despite the anisotropy of the aquifer created by preferential fracturing parallel to the fault-zone, fractures of all orientations contribute in varying degrees to recharge and movement of groundwater within the aquifer (Wermund, 1976; Wermund and Cepeda, 1977a,b).

Subdivisions of the Aquifer. Maclay and Small (1976, p. 22-25, 1975, 1983) subdivided the Edwards Aquifer into eight informal units within the San Marcos Platform in Medina, Bexar, Comal, and Hays Counties on the basis of subsurficial data such as test-hole-cores and geophysical logs. This scheme is summarized in Table 6 and applies to the zone of fresh water of the aquifer in the vicinity of San Antonio where it is generally confined at the base by the Glen Rose Formation and at the top by the Del Rio Clay.

Porosity. The development of porosity in the rocks of the Edwards Aquifer depended largely on the texture and structure of the carbonate facies and on the chemical and mineralogic properties of the rock and groundwater. These effects have been identified by Maclay and Small (1976, p. 25-30) and are briefly reviewed here.

Sedimentary structure such as burrows and solution-breccias contributed significantly to development of porosity in some carbonate facies. Some intertidal, burrowed beds of mudstone were partially dolomitized by circulating brackish water. Later, in some facies meteoric waters selectively dissolved the dolomitic fillings of burrows, leaving behind a very porous, honeycombed rock. This is characteristic of the Kainer Formation. Elsewhere, in other environments, fillings of burrows have been cemented or recrystallized to a tight mosaic of dolomitic crystals.

Intraformational brecciation, associated with evaporite rocks in supratidal deposits, has also contributed to porosity, particularly in the Kainer and Person Formations. Pores in some of these units are vugular and well connected.

Molds of allochems, especially those resulting from dissolution of fossils, are very common in the Edwards Aquifer. Locally, honey-comb rock may be present where molds of reef-forming organism are plentiful. This is generally true in the upper part of the aquifer.

Intraparticle, interparticle, intercrystalline, and fracture-formed porosity are common in the aquifer are common in the aquifer. Vugular spaces, which range in size from pinpoint up to several millimeters in diameter, can increase effective porosity in zones where the other types of porosity also exist. Fractures by themselves contribute little to total porosity; however, they generally increase the effective porosity and permeability by connecting zones of different types of pores.

By comparing stratigraphically equivalent rocks of the zones of fresh and saline water, Maclay and Small (1976, p. 27) were able to observe the effect of early-formed porosity on development of later-formed porosity. Their findings show that (1) dolomitic rocks that were highly vugular and porous in the saline zone are correlative with cavernous rocks in the zone of the fresh water, and (2) some less permeable, dolomitic rocks within the saline zone are represented by dense, recrystallized limestone in the zone of fresh water.

Diagenetic alterations have modified the porosity of rocks of the Edwards Aquifer, including limestone, dolostone, and mudstone. Dolomitization generally increased the effective porosity within the saline zone. In some places, dedolomitization—recrystallization of dolomite to calcite or spar—has produced tightly interlocking mosaics of calcitic crystals that commonly result in very dense, nonporous rock. Selective leaching of fossils has produces moldic porosity. Indiscriminant leaching along some fractures has also increased porosity. Conversely, cementation and replacement have greatly reduced porosity elsewhere.

Porosity may have been substantially enhanced by dissolution of specific minerals. The original sediments of the Edwards Aquifer included aragonite, gypsum and anhydrite. Aragonite was readily converted to stable calcite in waters of low salinity during periods of emergence; and fossils of molluscs, which preferentially precipitated aragonite in their shells, were selectively leached by marine and fresh water. Evaporites were periodically leached by meteoric waters during times of emergence in the Cretaceous Period. This allowed intraformational collapse and brecciation and a concomitant increase in porosity.

Dissolution and precipitation are occurring simultaneously throughout the entire vertical section of the aquifer in the zone of fresh water, as shown by a study of cores from test holes (Maclay and Small, 1976, p. 30). Crystals of calcite formed by precipitation have been observed on surfaces of fractures formed by dissolution.

Other aspects of chemical processes may have had a significant effect on the development of porosity of the Edwards Aquifer, but these are presently poorly understood. They include rates of reaction, mixing of chemically dissimilar waters, and changes in the temperature of groundwater. Abbott (1974, 1977b), Longman and Mench (1978), Mench (1978, written communication), and Woodruff and Abbott

(1979a) have recently studied these and other possibilities.

Bad-Water Line. The bad-water line, which separates the zones of fresh and saline water within the Edwards Aquifer, is designated as the trace of the contour of 1000 mg/l total dissolved solids (Figures 11, 12, and 13). This line is most accurately mapped in Bexar, Comal, and Hays Counties, and less accurately in Uvalde and Medina Counties owing to a larger number of sampled wells in the former. Lithology and chemistry of water differ greatly between the zones of fresh and saline water (Maclay and Small, 1976, p. 30-31).

In the zone of fresh water rocks are commonly recrystallized, light colored, have secondary and fracture-formed porosity, and contain little or no organic material. In the saline zone rocks are typically dark gray to dark brown, have textually controlled porosity, and contain unoxidized material. The dolomite-to-limestone ratio is much lower and the fabric of the carbonates is much more altered in the zone of fresh water than in the zone of saline water.

Water within the zone of fresh water has a considerably lower concentration of dissolved ions than that in the saline zone. The oxidation potential is positive (oxidizing) in the zone of fresh water and negative (reducing) in the saline zone. The pH of water in the zone of fresh water generally ranges from 7.0 to 7.6, and it is commonly less than 7.0 in the saline zone (Maclay and Small, 1976, p. 32). Tritium and carbon-14 studies indicate that water in the zone of fresh water is much younger than water in the saline zone (Pearson and others, 1975).

Several possible explanations for the position of the bad-water line have been given (Maclay and Small, 1976, p. 32). Fresh water may not have circulated through the saline zone for prolonged periods of time, as suggested by the occurrences of organic materials and less altered textures within the rocks of that zone. Faults control the location of the major springs, and this in turn may have affected the position of the line. A series of captures of springs at high elevation by a dominant spring at a lower elevation could explain a migration of the bad-water line to a lower elevation (Maclay and Rettman, 1972). In places the position of the bad-water link may also result from an increase in pressure from overburden that has locally reduced both width of fractures and permeability.

Relationships between Precipitation and Recharge. Puente (1975) studies the correlation of recharge of groundwater to precipitation for the Edwards Aquifer. Annual recharge obtained from historical records of streamflow and mean-annual precipitation computed from records of rainfall were used to develop simple equations by linear-regression for seven sub-basins in the San Antonio area. The equations of modeling, which were derived for use in eliminating annual recharge to the aquifer, are of the form $R = CP^a$, where R is the annual recharge in acre-feet (millions of cubic meters), P is the adjusted weight, mean-annual precipitation in inches (millimeters), C is the regression constant, and a is the regression coefficient. This method is inherently crude, as the standard errors of estimate of these equations ranged from 26 percent for the Blanco River basin to 45 percent for the areas between the basins of the Sabinal and Medina Rivers.

Figure 15 shows the relationship between annual recharge and precipitation for the seven subbasins. Puente (1975) computed estimated, annual recharge using the equations of regression and compared this with estimates made on the basis of observed streamflow and precipitation. Computed and observed curves of recharge for the period 1954 to 1970 compared favorably. The derived equations may thus be used to make gross estimates of annual recharge from adjusted, weighted, mean-annual precipitation.

Relationships between Recharge and Discharge. Approximately 35 percent of the total recharge to and about 90 percent of the total discharge from the Edwards Aquifer occurs in Bexar, Comal, and Hays Counties (Puente, 1976). Five of the six largest springs draining the Edwards Aquifer are in these counties, and their combined, annual discharge was about 60 percent of the annual recharge to the aquifer for the period 1945 through 1973 (Puente, 1976).

Puente (1976) performed a statistical analysis on data representing waterlevels, springflow, and streamflow in the San Antonio area. By analyzing the hydrogeologic setting of major springs in Comal and Hays Counties, he found that large springs in the area of San Antonio occur along faults that serve as natural outlets for discharge of water from the Edwards Aquifer. The major springs are the Leona River Springs near Uvalde, San Antonio and San Pedro Springs in San Antonio, Comal and Bexar Springs at and near New Braunfels, and San Marcos Springs at San Marcos (see Figure 12 for location). Figure 16 shows the hydrographs of Comal, Hueco, and San Marcos Springs from 1934 through 1973. Additional data is available in Guyton and Associates (1979).

Comal Springs (Figure 14) have the greatest discharge of any known springs in Texas and in the southwestern United States, with an average flow of 9.2 m^3/s. Water rises along a 640 m-long line at the base of the escarpment formed by the Comal Springs Fault. The throw of the fault at this location is approximately 150 m. Less permeable, Late Cretaceous rocks are faulted down against the cavernous Edwards Group. This forces groundwater to rise to the surface at artesian springs along the traces of the fault (George and others, 1947, 1952; Eisele and Gray, 1973; Brune, 1975; Guyton and Associates, 1979).

The waters from the spring are crystal-clear and maintain a nearly constant temperature of about 23.3° C, which is 3.3° higher than the average air-temperature at New Braunfels. This and a fairly constant chemical composition of water from the spring suggest that migrating waters in the reservoir may circulate at depths as great as 90 to 150 m below the surface. Hydrographs of wells west of Comal Springs show a strong correlation between flow from Comal Springs and fluctuations in wells west of Comal Springs (Puente, 1976). Fluctuations in the discharge of Comal Springs apparently reflect changes to the rate of pumping in the area of high withdrawal in Bexar County. This relationship is consistent with the knowledge that most discharge of Comal Springs is derived from underflow from the artesian zone west of Comal Springs (Puente, 1976).

The only recorded cessation of flow from Comal Springs occurred from 13 June

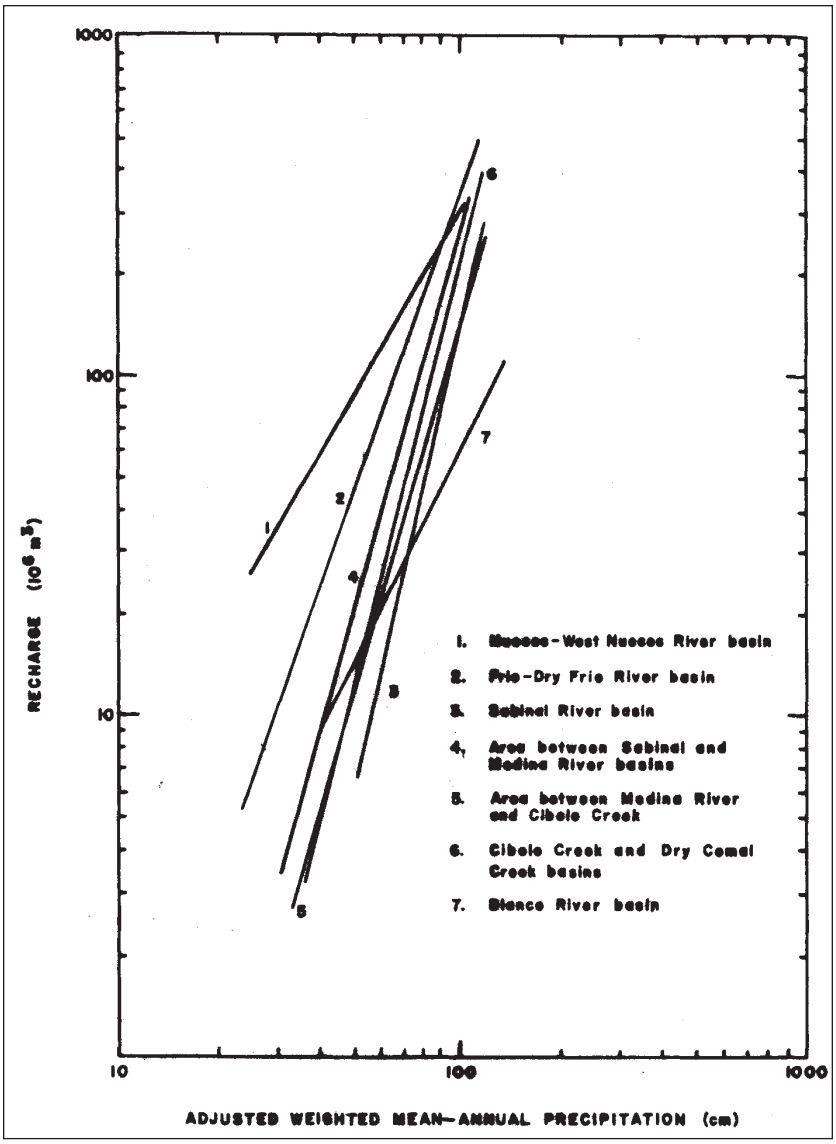


Figure 15: Relationship between annual recharge and precipitation for seven sub-basins in the area of San Antonio. Modified from Puente (1975, figures 5-11, p. 13-19).

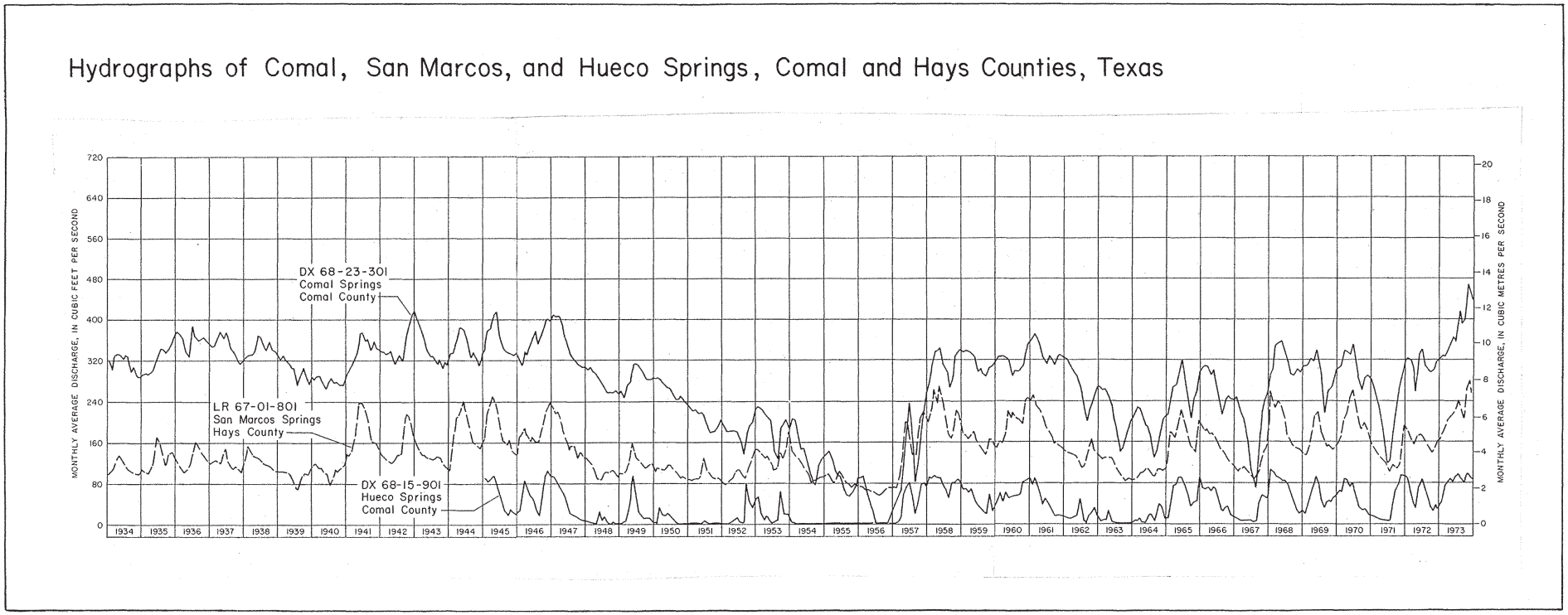


Figure 16: Hydrographs of Comal, Hueco, and San Marcos Springs. Modified from Puente (1976, figure 6, p. 21). See also Guyton and Associates (1979, figure 13, p. 43).

to 4 November 1956 (Figure 16). This was caused by a severe drought coupled with increased pumpage in Bexar County Waterlevels dropped to the altitude of the outlet of the spring at 189.9 m above mean sealevel.

Hueco Springs, along the Guadalupe River approximately 5 km north of New Braunfels, issue from two outlets within a faulted zone along the Hueco Springs Fault. The lowest outlet, at an altitude of approximately 198.7 m above mean sealevel, lies about 8.8 m higher than the outlets of Comal Springs. Hydrographs of Hueco and San Marcos Springs are generally similar to those of Comal Springs, but occasionally the trends of discharge differ significantly (Figure 16).

Studies of Hueco Spring (George and others, 1947, 1952; Petitt and George, 1956; Guyton and Associates, 1958, 1979) determined that the area of recharge for the springs encompasses the drainage-basins of Dry Comal Creek and other tributaries to the Guadalupe River north of the Hueco Springs Fault and west of the river. Recorded Waterlevels in the area northwest of Hueco Springs indicate that the hydraulic gradient is sufficient to provide flow to the springs.

Fluctuations of waterlevels in wells just north of the Hueco Springs Fault correlate best with variations in discharge of Hueco Springs. The area of recharge for these springs is relatively small. This is demonstrated by (1) rapidly rising springflow for short periods following heavy precipitation in the area, (2) variations of as much as several degrees in the temperature of water issuing from the springs (the temperature is not as constant as that of water from Comal Spring), and (3) a high concentration of tritium in water from the springs, indicating that the water was recently recharged to the aquifer (Puente, 1976). Hueco Springs frequently cease to flow for several months during extended periods of low precipitation in the vicinity (Figure 16).

Discharge from San Marcos Springs occurs at an altitude of 175.0 m above mean sealevel or approximately 14.9 m lower than the outlets at Comal Springs in New Braunfels. San Marcos Springs are along or very near the San Marcos Springs Fault where it comes near to the Comal Springs Fault (Guyton and Associates, 1979).

Fluctuations in discharge of San Marcos Springs and Hueco Springs are similar. This suggests that a significant part of the flow from the former is derived from recharge entering the aquifer in Comal and Hays Counties north of the Hueco Springs Fault. Moreover, hydrographs of waterlevels in wells east of Comal Springs and of discharge from San Marcos Springs have similar patterns of fluctuation, suggesting that the wells and San Marcos Springs are in hydraulic continuity. Furthermore, this indicates that recharge occurs in the area between Comal Springs and San Marcos Springs (Puente, 1976).

During the period of drought, which began in 1947 but was severe from October 1955 to February 1957, Hueco and Comal Springs recorded little or no flow, and streams in the area provided little recharge to the aquifer (Petitt, 1963). During this time, springflow at San Marcos Springs was sustained by underflow from the artesian part of the aquifer to the west. Puente (1976) and Guyton and Associates (1979) interpret that this data indicates that discharge from San Marcos Springs is derived, in part, from regional underflow from the area of Comal Springs to the southwest and, in part, from local recharge from northern Comal County and Hays County. The discharge of San Marcos Springs during periods of normal or above-normal recharge in the vicinity is composed of both underflow and local recharge. Under these conditions fluctuations in discharge from San Marcos Springs differs significantly from those in discharge of Comal Springs owing to the imposed effects of local recharge (Puente, 1976). Many of these findings are corroborated by modeling of the Edwards Aquifer by Klemt and others (1975, 1979).

Concentrations of tritium measured in the groundwaters of the Edwards Aquifer corroborate many of the findings state above (Pearson and others, 1975; Rettman and Pearson, 1975). These data indicate that much of the water discharged from Hueco and San Marcos Springs is derived from recharge in northern Comal County and Hays County. This recharge does not mix with water from the deeper, artesian part of the Edwards Aquifer.

In partial summary: (1) flow to Comal Springs is mostly regional underflow moving through the deeper part of the aquifer adjacent to the Comal Springs Fault in Comal County, (2) Hueco Springs are fed mainly from local recharge in the area of drainage northwest of the Hueco Springs Fault and west of the Guadalupe River in Comal County, and (3) San Marcos Springs are fed by regional underflow from the area of Comal Springs and from local recharge in northern Comal County and Hays County (Puente, 1976, p. 57).

Geochemical Studies of the Edwards Aquifer. Pearson and Rettman (1976) studied the chemistry of waters of the Edwards Aquifer. Analyses included 14 common chemical species, and stable carbon, radiocarbon, and isotopes of sulfur. Pearson and Rettman were able to group waters of the aquifer into five chemical types: (1) recharged water, (2) main, fresh water, (3) varied water, and (4) two types of transitional and saline water. All types were found to be saturated with respect to calcite, and all transitional and saline waters were saturated with respect to dolomite as well. Some saline samples of water were also saturated with respect to gypsum, celestite, strontianite, and fluorite.

Abbott (1977b) indicated that undersaturated waters exist at least seasonally in the eastern part of the Edwards Aquifer, despite distances of travel for some of the groundwater exceeding 200 km within the carbonate rocks. Abbott (1977b) and Woodruff and Abbott (1979a) attributed this anomaly, in part, to large volumes of groundwater flowing in pipe-like voids where little of the water is actually in contact with the host-rock, and, in part, to the mixing of karstic water. Undersaturation may occur if recharge to the main, saturated body of groundwater is cooled upon entering or if it is mixed with saturated water of different calcitic concentration or of different partial pressure of carbon-dioxide. This effect was originally proposed by Bogli (1964, 1971) and later discussed and evaluated by Thrailkill (1968), who showed that it can be insignificant under some conditions.

R.D. Reeves (1976) studied the chemical and bacteriological quality of groundwater from the aquifer and reported that concentrations of dissolved solids ranged from 200 to 527 mg/l. In general, the waters are very hard (greater than 180 mg/l as calcium carbonate) and are of the calcium-bicarbonate-type. Other chemical studies include those of Rightmire and others (1974), who investigated the sources of nitrates in the Edwards Aquifer, and St. Clair (1979), who investigated the quality of water in the Edwards Aquifer in Travis County.

Evolution of the Edwards Aquifer. Woodruff and Abbott (1979a,b) have proposed a developmental model of the Edwards Aquifer which relates the integration of paths of flow of groundwater to the evolution of drainage-basins on the Edwards Plateau. Their hypothesis is summarized as follows:

1. The San Marcos Arch was exposed to subaerial erosion during the late Early Cretaceous. Porosity was enhanced in the eastern part of the area of study by increased circulation of groundwater and accompanying dissolution of the carbonate rocks. At this time the loci of greatest discharge were located in this area.

2. The greatest displacement during Balcones faulting in the Miocene Epoch occurred in this same area, and fractures associated with faulting provided additional porosity, permeability, and avenues for circulation of groundwater.

3. Vertical displacements owing to faulting resulted in increases in hydraulic gradients and circulation of groundwater along the Balcones Escarpment. The trend of faults was at an acute angle to the courses of main streams of the ancestral Guadalupe and San Antonio River Systems. Surficial streams normal to the escarpment began to incise rapidly into northwesterly trending canyons.

4. These streams of steep gradient captured eastward-flowing, major streams in the eastern watersheds of the Edwards Plateau.

5. Pirated Stream incised into the aquifer at the lowest topographic level within the region. This was brought about by the sudden acquisition of extensive areas of catchment in a subhumid area, and by steep gradients of streams that reflected larger displacement along faults in the eastern part of the Plateau.

6. The low topographic points of discharge became the loci of the major springs that now drain the aquifer.

7. Recharge became dominant in the basin of the Nueces River in the western part of the area because here stream crossed permeable limestone at higher topographic elevations than in the basins of the Guadalupe and San Antonio Rivers. Other factors included lesser displacements along faults, no major piracy of streams, and less vigorous erosion under semi-arid, climatic conditions.

Trinity Group Aquifer

The Trinity Group Aquifer of north-central Texas extends into the northeastern part of the area of study (Figure 4). Water is found in units of sandstone of the Trinity Group. These units generally do not yield large quantities of water to wells; however, they are important sources of water for small communities where they supply potable water to wells of small yield (Holloway, 1961; Mount and others, 1967). The Cow Creek Limestone has much higher yields and locally is an important component of the aquifer.

Minor Aquifers

Two significant minor aquifers occur in the area of study. Both are found in the Llano basin (Figure 5).

Ellenburger-San Saba Aquifer

Although the Ellenburger Group and San Saba Member of the Wilberns Formation are separate geologic units, they are difficult to distinguish hydrologically and

are therefore considered as a single hydrostratigraphic unit. Water in this aquifer occurs in vugular, or cavernous, zones an din fractures and joints that, in places, have been enlarged by dissolution. Most units of the aquifer yield water; however, the dolomitic facies of the Staendebach Member is generally most favorable for supplying water to wells (Cloud and Barnes, 1946a, p. 127). Water-bearing zones are not horizontally continuous over large areas and are generally under artesian pressure.

Recharge to the aquifer is derived from precipitation falling on permeable outcrops on the uplands and from streams flowing over permeable rocks of the aquifer. Discharge is generally from springs along surficial streams. Baseflows of many streams in the Llano Basin are supported by discharge from this aquifer.

Water occurs in solutional openings, but these vary in size and degree of interconnection. Large yields are common from wells located in areas of extensive, solutional permeability. Additional data on chemical quality, waterlevels, and availability of groundwater are given my Mount and others (1967, p. 71-75).

Hickory Aquifer

The Hickory Sandstone Member of the Riley Formation is an important aquifer north and west of the Llano Basin. The lower part of the aquifer yields water more rapidly than does the upper part. Downdip, where the Hickory Sandstone is overlain by other rocks, the aquifer is artesian. Waterlevels in many updip areas have been significantly lowered by irrigation, causing problems of depletion in some places.

Recharge is derived from precipitation and infiltration from vertical fractures within the overlying Cap Mountain Limestone. Discharge is through seeps along surficial streams. Further information on chemical quality, waterlevels, and availability of groundwater can be found in Mount and others (1967, p. 75-79).

Other Aquifers

Other rocks of the area of study yield water, but their importance as aquifers and to the investigation of karst is minimal. These rocks include most Precambrian and Paleozoic units, Cretaceous rocks younger than the Edwards Group, and alluvial deposits. Some of these are discussed by Mount and others (1967, p. 79-80).

ADDITIONAL INFORMATION

Numerous studies of surficial water and groundwater have been undertaken within areas ranging in size from major drainage-basins and counties (Table 5) down to local sites, such as specific municipal water-wells. References to much of the pertinent literature are in the bibliographies of Friebele (1975) and Friebele and Wolff (1976). Additional references on the hydrogeology of the Edwards Plateau are listed in Metcalf and Eddy, Inc. (1976).

PART II

INVESTIGATIONS
IN SELECTED AREAS

4
DISTRIBUTION OF KARST AND CAVES

KARSTIC REGIONS OF TEXAS

Texas has several important, karstic regions (Table 7; Figure 17). These have been identified and delineated by the Texas Speleological Survey based on the characteristics and distribution of caves (Russell, 1960, 1968; Reddell and Russell, 1962; Reddell, 1964b; Smith, 1965, 1968, 1969, 1971a; Fieseler and others, 1978). Eight of the thirteen regions of Figure 17 lie largely within the area of study and account for approximately 80 percent of all recorded caves in Texas (based on 1970 tabulations).

A comparison of Figure 17 with Figure 2 shows that karstic regions correspond closely to physiographic divisions. This is as expected because the geologic setting has guided development of both the landscape and the caves within each particular division.

Other attributes besides caves are important in characterizing karstic regions. These include the character and distribution of surficial, karstic landforms, the dominant processes and controls that have acted to produce karst, and the relationship of karst to the regional, hydrogeologic setting.

To better define the karstic regions of the Edwards Plateau I have selected specific areas for detailed analysis. By comparing these areas it is possible to isolate factors that either have universally influenced karstification in the region or have dominated only in specific locations.

DISTRIBUTION OF SURFICIAL KARST ON THE EDWARDS PLATEAU

Nearly all areas of the Edwards Plateau that are underlain by carbonate rocks exhibit karstic landforms. These range in size from minor solutional sculpture, such as pitting and rills, to large-scale features that include solutional valleys and regional zones of collapse. Most exposures of carbonate bedrock are karsted.

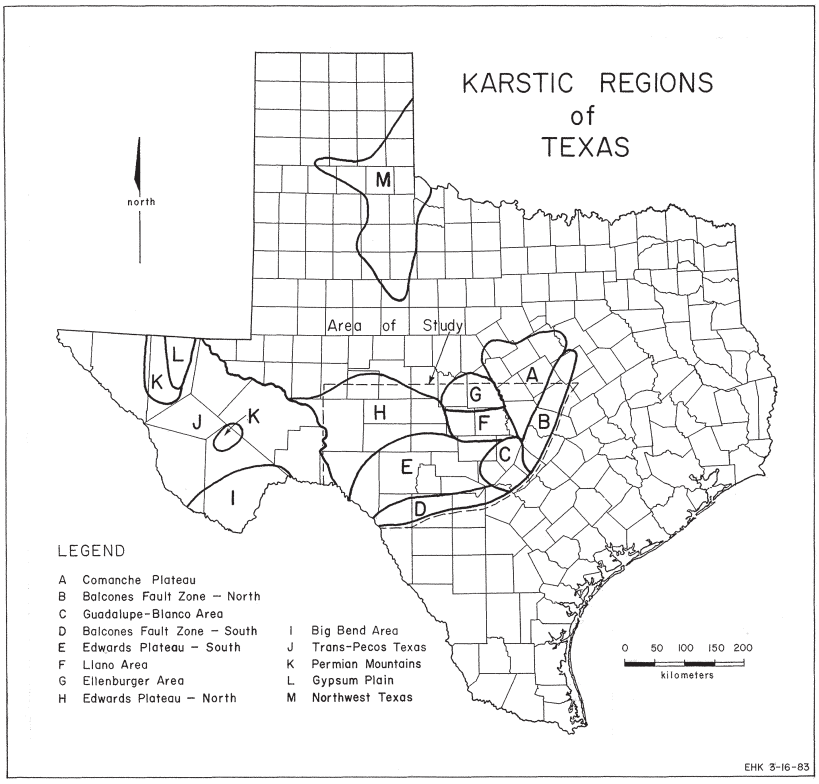


Figure 17: Karstic regions of Texas. Boundaries and names of karstic regions are modified from Fieseler and others (1978, p. 19).

DISTRIBUTION OF CAVES

The distribution of known caves on the Edwards Plateau is presented in Appendix C and Plate 1. It is readily apparent from these compilations and Table 7 that caves are not distributed uniformly throughout the area. A greater concentration of known caves occurs in the eastern part of the area, particularly along the Balcones Fault Zone and in the northern part of the Llano Basin. Caves are sparse in the northwest part of the area of study and within the Lampasas Cut Plain (Comanche Plateau). The general increase in numbers of caves from west to east can, in part, be attributed to geologic factors (Woodruff and Abbott, 1979a,b).

However, exploration and surveying of caves have been more intense in the eastern part of the Plateau because this area is closer to metropolitan areas and many roads provide easy access to caves. Consequently, knowledge of the distribution of caves may be substantially biased by the level of effort areas have received from explorers. Nevertheless, some aspects of the distribution of caves, such as that of the relatively long and deep caves (Appendix C), are useful for regional analysis of karstification.

SELECTED AREAS FOR DETAILED INVESTIGATION

The large size of the area of study and the time allotted for completion of research made it necessary to distribute fieldwork in such a way that significant localities would be adequately investigated. Eight subareas within the overall area of study were selected for detailed analysis and are outlined on Plate 1. These were chosen for several reasons: (1) they contain significant karstic features and important caves, (2) they are distributed geographically over the Edwards Plateau and thereby reflect diverse geologic, hydrologic, climatic, and speleologic settings, and (3) a substantial amount of basic data is available for each, including geologic maps, hydrologic information, and surveys of caves. The specific areas of investigation and the importance of each are summarized in Table 8.

The occurrence and origin of pseudokarst developed on granite in the Llano Basin was also investigated. Although this is somewhat incidental to the main thrust of the research, it nevertheless provides additional information on the geomorphic evolution of the Edwards Plateau.

The following chapters of Part II discuss the detailed investigations of the eight specific areas of carbonate karst (Chapters 5 through 12), and that on granitic karst (Chapter 13). Each chapter includes a synthesis and summary of the findings on that particular area. A regional synthesis and summary is presented in Chapter 14.

Table 7: Karstic Regions of Texas[a]

Karstic Region[b]	Approx. Area (km^2)	Bedrock	Physiography
		Carbonate Areas	
Comanche Plateau	9000	Edwards Group	Dissected plateau, rolling plains
Balcones Fault Zone - North	6900	Edwards Group, Glen Rose Formation	Escarpment, rolling
Guadalupe-Blanco Area	2300	Glen Rose Formation	Deeply dissected, incised
Balcones Fault Zone - South	7000	Edwards Group,Glen Rose Formation	Escarpment, rolling
Edwards Plateau -South	19400	Glen Rose Formation, Edwards Group	Deeply dissected
Llano Area	5100	Paleozoic and Precambrian rocks	Dissected plateau, rolling plains
Ellenburger Area	4900	Ellenburger Group, other Paleozoic Rocks	Dissected plateau, rolling plains
Edwards Plateau - North	21400	Edwards Group	Incised plateau, rolling plains
Big Bend	2800	Cretaceous rocks	Mountainous
Trans Pecos	77000	Mesozoid and Cenozoic Rocks	Basin and range, rolling prairie
Permian Mountains	2300	Permian rocks	Mountainous
		Gypsum Areas	
Gypsum Plain	1300	Castille Formation (Permian)	Low rolling plains
NortheWest Texas	24100	Blaine Formation (Permian)	Low rolling plains, dissected

[a]Modified from Smith (1971a, p. 3).

[b]Names of karstic regions are retained from those of the Texas Speleological Survey (Smith, 1971a; Fiesler and others, 1978; see also Table 4.

Table 8: Selected Areas for Detailed Investigations

Langtry Area, Val Verde County (Chapter 1):
Some of the deepest caves of Texas were developed in the Devils River Formation, a complex of reefal and interrefal rocks. Important caves include Emerald Sink, Fisher's Fissure, Langtry Gypsum Cave, Langtry Lead Cave, Langtry Quarry Cave, and others. Large areas of solutional collapse represent extensive, regional kartification in the past.

Sonora Area, Sutton County (Chapter 6):
Large, multi-level caves such as Caverns of Sonora and Felton Cave have formed here. In many ways this area is typical of the semi-arid, western part of the area.

Menard Area, Menard County (Chapter 7):
This is the site of the Powell's Cave System, the longest cave surveyed in Texas (in excess of 17 km mapped). Nearby Silvermine Cave is hydrologically related to Powell's Cave. This system has experienced several distinct episodes during its development.

Bend Area, San Saba and Lampassas Counties (Chapter 8):
This location is the site of several long caves that are genetically related to the incision of the Colorado River. Long caves, such as Gorman Cave, Clarks Branch Well Cave, and others, have formed in rocks of the Ellenburger Group. Karst of this area is locally highly fissured and the greatest known density of distribution of caves occurs in this part of the area of study.

Burnet Area, Burnet County (Chapter 9):
As in the Bend area caves at this locality have developin in rocks of the Ellenburger Group. However, they are generally located well above stream at baselevel. Longhoen Cavern, a Texas State Park, is the best example.

New Braunfels Area, Comal and Bexar Counties (Chapter 10):
This area is located within the Balcones Fault Zone and many caves are intimiately related to recharge of the Edwards Aquifer, particularly along Cibolo Creek. Several large caves, such as Natural Bridge Caverns, Bracken Bat Cave, and Double Decker Cave are located on drainage divides near the Bat Cave Fault.

Austin-San Marcos Area, Travix, hays, and Comal Counties (Chapter 11):
The Austin-San Marcos area contains several long caves, such as Airmans Cave, Cave X, and Wonder Cave, and numerous small caves. Extensive fracturing related to Balcones Faulting has greatly controlled development of caves.

Georgetown Area, Williamson County (Chapter 12):
Caves are generally small in the northeast corner of the area of study. However, Inner Space Coverns, one of the longest caves in Texas (over 4.6 km mapped), is found here. The area is within the Balcones Fault Zone.

5
LANGTRY AREA

DESCRIPTION OF AREA

The Langtry area includes the Langtry and the southern part of the Harkell Canyon 7.5-minute U.S.G.S. quadrangles, located in Val Verde County (Plate 1). This, the southwesternmost subarea investigated in this study, is the only one west of the Pecos River and within the Stockton Plateau division of the Edwards Plateau physiographic section (Figure 2). The area is drained by the Rio Grande to the south and by its tributary, the Pecos River, to the north. Some of the deepest caves in Texas are in this locality (Appendix C).

Roberts and Nast (1918), International Boundary and Water Commission (1950-1951), Freeman (1964a, 1968), Brown and others (1965), Reeves and Small (1973), and Barnes (1977) have mapped the geology of the area at scales ranging from 1:50,500 to 1:506,900. The geologic map of the area (Figure 18) has been modified from Freeman (1964a) to show currently accepted stratigraphic nomenclature of Lozo and Smith (1964), Rose (1968, 1972), and Barnes (1977).

Darton (1933) described the geology along the Southern Pacific railroad in the vicinity of Langtry. The geology of adjacent and nearby areas in Val Verde and Terrell Counties has been mapped at scales of 1:62,500 (Sharps, 1963, 1964; Freeman, 1964b,c, 1965, 1968; Sharps and Freeman, 1965) and 1:31,680 (Spice, 1954).

The Langtry area is underlain by Lower and Upper Cretaceous rocks. Local deposits of Pleistocene gravels are found at sites of former terraces in the vicinity of the Rio Grande and Pecos River, and Holocene alluvium floors the valleys of the Rio Grande and Langtry Creek (Figure 18).

The lowest exposed unit of bedrock is the Devils River Formation. It is a light-gray to yellowish-gray, thin- to very thick-bedded limestone or mudstone which is finely crystalline and very fossiliferous. It is approximately 260 m thick in the Langtry area (Reeves and Small, 1973, figure 10, p. 21), but only the top 60 m are exposed by erosion near the Rio Grande and Pecos River.

The Devils River is overlain by the Buda Limestone, which is a light yellowish-gray to pale-orange, thick- to very thick-bedded, dense, porcelaneous (micritic) limestone in its upper and lower part, and a yellowish-gray, thin- to thick-bedded, clayey limestone with nodular weathering in its middle part. The Buda is 24 to 27 m thick in the area and exposed along slopes of valleys of the Rio Grande and Pecos River.

The Boquillas Flags overlie the Buda limestone and consist of four distinguishable units of dark-gray, olive-gray, brownish-gray, and yellowish-gray, clastic, clayey, silty limestone. Beds of the basal unit are medium-grained, cross-laminated, calcarenitic beds and pinch and swell along the strike. Beds of the next overlying unit are in part evenly laminated and in part flaggy and commonly contain very thin, bentonitic beds. Above this is a unit of thin-bedded, resistant limestone that weathers to light-gray ledges and talus. Beds of the uppermost unit are poorly resistant, laminated, and weather to slopes free of coarse debris. The Boquillas Flags are approximately 47 to 55 m thick in the area and generally thin toward the south. Of all the formations in the region this one is the most extensive in outcrop and is exposed along the divide between the Rio Grande and Pecos River.

The youngest unit of bedrock in the area, which overlies the Boquillas Flags, consists of very pale orange, very fine-grained limestone in thin to thick, structureless beds with very thin to thin interbeds of rusty-colored friable, laminated, clayey limestone. The original thickness of the unit in this vicinity is unknown, but up to 45 m are exposed where the formation caps ridges in the divide between the Rio Grande and Pecos River (Freeman, 1964a). This unit has been mapped as Austin Chalk by Freeman (1964a) and Barnes (1977), but the type-locality of the Boquillas Flags contains Austin Chalk equivalents. Thus, the designation Austin Chalk is considered to be erroneous (K. Young, written communication) and the unit should be considered an upper part of the Boquillas Flags. However, the Austin Chalk designation is retained here for consistency with the published maps.

The Cretaceous strata strike N 45°-90° W and dip gently to the southwest at 0.3° to 1.0° (Calvert, 1928; Freeman,1964a; Rees and Buckner, 1980, figure 2, p. 5). The dip is greatest in the northeast and near the Rio Grande and least in the center of the area. This is shown in the geologic section of Figure 19.

Calvert (1928) described the geologic structure of Val Verde County, including the southwest regional dip and fault zones. Flawn (1959a,b), Flawn and others (1961), and Cunningham (1961) studied structural elements of the Ouachita System, the Val Verde Basin, and the Devils River Uplift, all of which underlie the area. Leonard (1977a,b, 1978) mapped joints, fractures, and lineaments in Val Verde County and related these to regional tectonism.

Calvert (1928) identified an anticlinal axis trending from Del Rio northwest to Comstock in Val Verde County. Although he did not name this structure, he noted that it had surficial expression in Cretaceous rocks of the area. Galley (1958) later identified an anticlinal uplift trending toward the south-southeast from the Marathon dome that is generally aligned and continuous with Calvert's structure to the southeast. Galley introduced the term "Devils River Uplift" for this feature.

The Devils River Uplift is an area of high-standing older rocks (presumed to be Precambrian) in the subsurface of southern Val Verde and Kinney Counties that parallels the Rio Grande (see Figure 10). Flawn (1959b, p. 74) remarks that the uplift "probably influenced the course of the Ouachita geosyncline in early Paleozoic time" and thereby represents an "ancient and persistent feature." Like the Llano Uplift, the Devils River Uplift was a positive feature throughout the Paleozoic Era, and it apparently continued to rise during this time (Webster, 1978, 1980, 1982; Webster and others, 1978). Uplift continued during the Cretaceous, as indicated by Cretaceous rocks that are warped accordingly. This probably explains the variations in attitude and thickness of beds shown in Figure 19 and on the geologic map by Freeman (1964a).

The positive expression of the Devils River Uplift created an open, shallow, marine environment with moderate to high wave-energy during deposition of the upper part of the Devils River Formation. This produced a thick sequence of reefal, bioclastic limestone of relatively uniform lithology along the trend of the uplift during the Lower Washita Substage (Rose, 1972, p. 52; Freeman, 1968). Chemical analysis of samples from the upper Devils River Formation collected in Val Verde, Kinney, and Real Counties show a high content of calcite (96.0 to 99.7 percent $CaCO_3$) (Rodda and Fisher, 1961; Rodda and others, 1966, p. 192, 230-234, 263-265; Freeman, 1968, table 1, p. 5). High purity, fine-grained texture, thick beds, and stratigraphic homogeneity have increased the solubility of this formation and thereby have promoted development of caves.

The Devils River Uplift has also contributed to development of fractures in the Langtry area. Leonard (1977a,b, 1978) showed that northwest-trending fractures coincide with the northern boundary of the Devils River Uplift and that they may have formed in response to the presence of the basal structure during compressional deformation along this trend in the Late Cretaceous or Early Tertiary.

Lineaments mapped by Freeman (1964a) have a dominantly northeastern orientation and correspond to northeast-trending sets of fractures and joints mapped by Leonard (1977a). Some fractures of this orientation are orthogonal to those of a northwestern orientation, suggesting that they too are related to the Devils River Uplift. However, Leonard (1977a) found a second set of northeast fractures that form a small dihedral angle with the first set. He suggested that uplift of the Edwards Plateau during the Late Cretaceous and Early Tertiary may have formed these fractures under tension.

Fowler (1956, 1957) indicated that tectonism in south-central Texas produced zones of compression and folding in the Rio Grande Embayment in the vicinity of the Devils River Uplift, and produced zones of extension and faulting in the northeast along the Balcones Fault Zone and San Marcos Arch. These were contemporaneous events during the Late Cretaceous and Early Tertiary. Leonard (1977a) suggested that the two northeast-trending, conjugate sets of fractures of

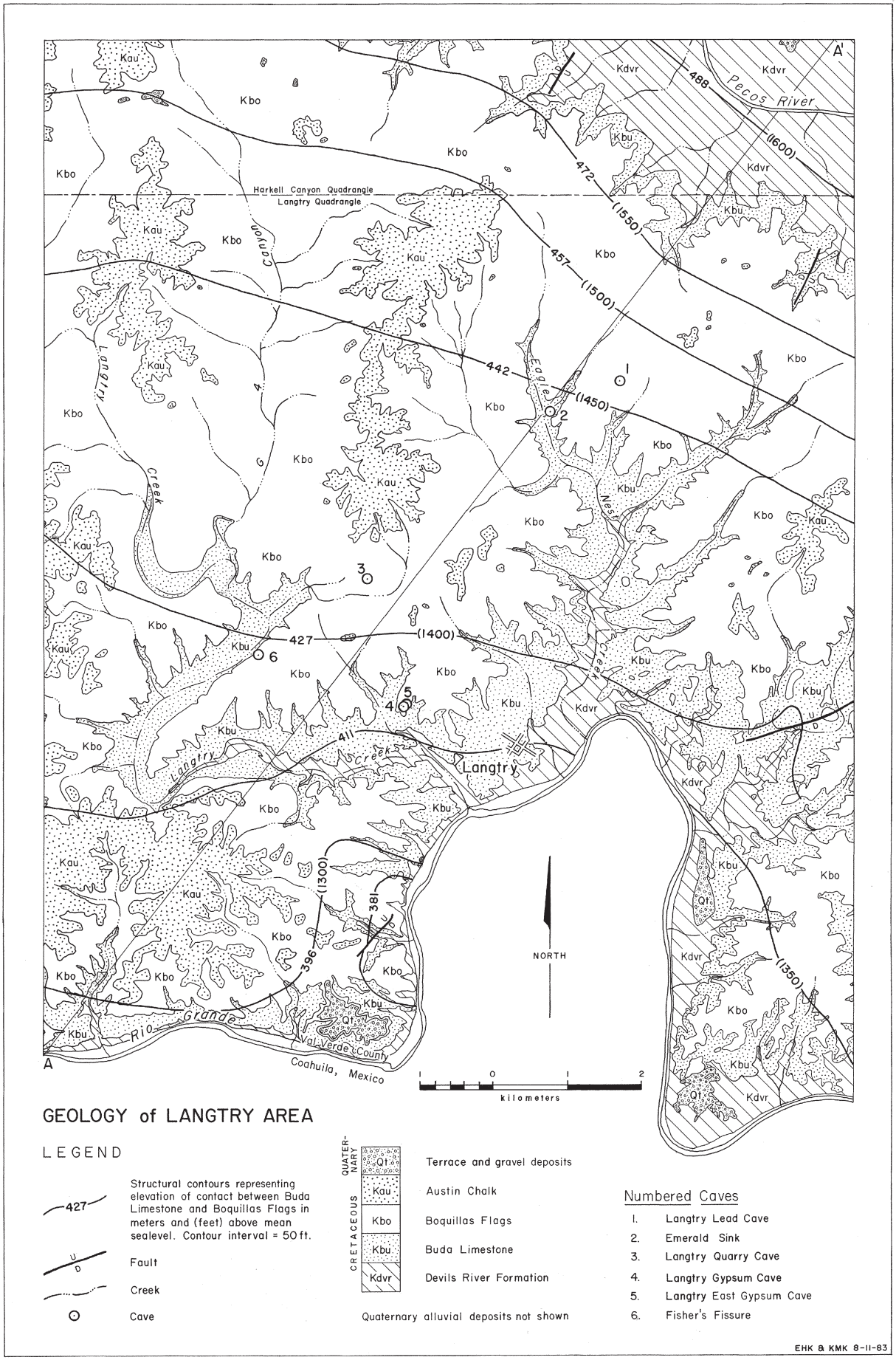


Figure 18: Geologic map of the Langtry area, Val Verde County. Modified from Freeman (1964a). Cross-section A-A' corresponds to Figure 19. Locations of investigated caves are shown.

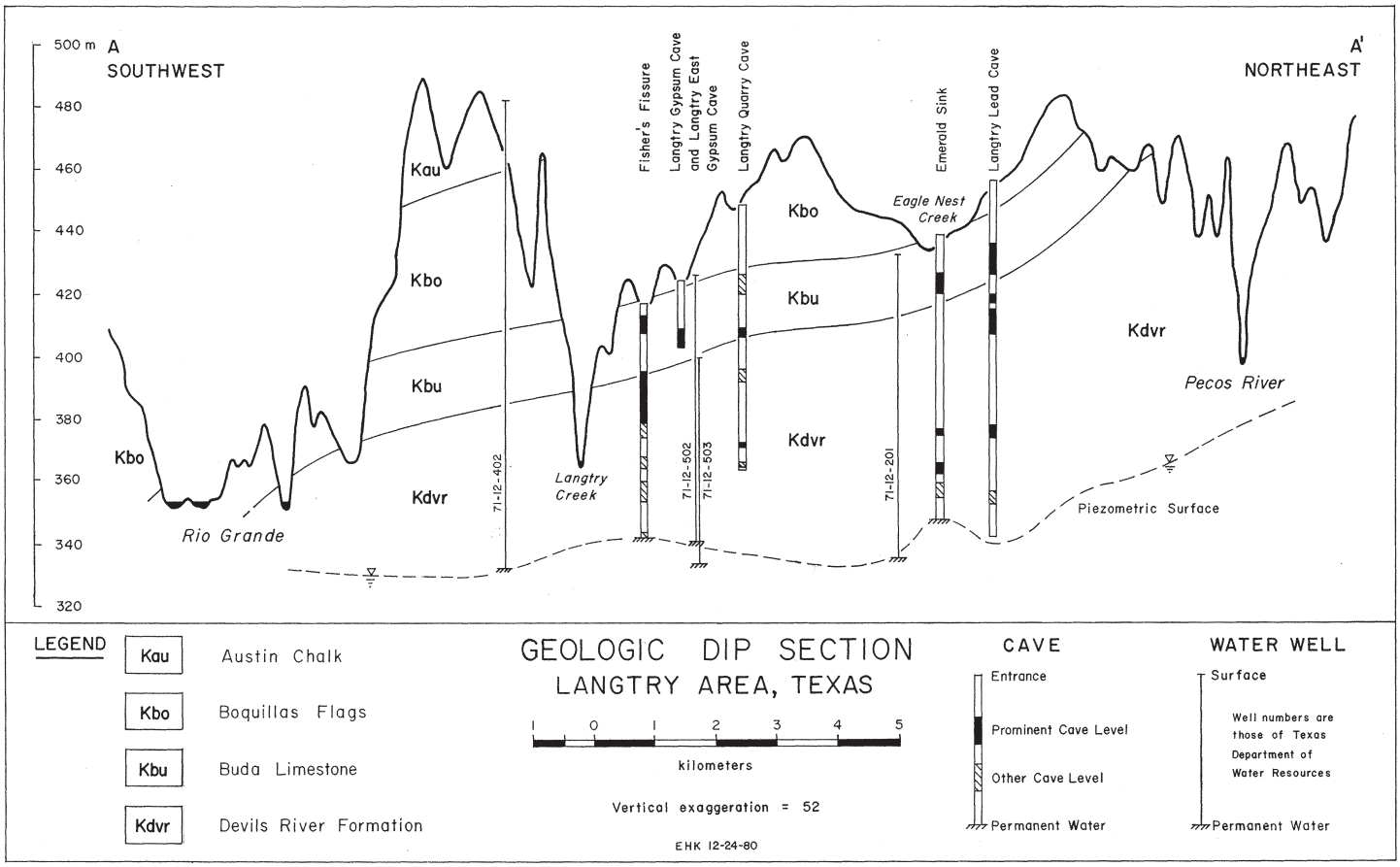


Figure 19: Geologic section along the regional dip in the vicinity of Langtry. See Figure 18 for location. Vertical extent and levels of passages of major caves are shown as projected onto plane of section. Modified from maps in Kunath and Smith (1968). Geology and topography are based on Freeman (1964a) and Barnes (1977). Piezometric surface and information from water-wells are from Brown and others (1965) and Reeves and Small (1973).

small dihedral angle may support Fowler's interpretation that conjugate sets of joints of small dihedral angle are associated with simultaneous compression and tension is also supported by findings of Parker (1942) and Muehlberger (1961).

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

Two caves of the Langtry area were known in 1948 (White, 1948b). Langtry Gypsum Cave and Langtry East Gypsum Cave were opened prior to this time during excavation for a highway. During the 1950's and early 1960's these and several other caves were explored and mapped by cave-exploration groups, leading to eventual documentation in the speleological literature (Reddell and Russell; 1961b, Reddell, 1963; Kunath and Smith, 1968).

The larger caves in the area are within 5.3 km of the town of Langtry and north of the large meander in the Rio Grande. Most entrances to caves are found in the outcrop of the Boquillas Flags and along the flanks of creek valleys (Figures 18 and 19).

Smith (1971b) and Reddell and Russell (1961b) briefly investigated the geology of caves in the area. They noted the stratigraphic position of levels within caves and correlated the orientations of passages in Emerald and Langtry Lead Caves with the orientations of photolineaments on the geologic map by Freeman (1964a).

An investigation of the hydrogeology along the Rio Grande from the Big Bend area east to Del Rio by the International Boundary and Water Commission showed that the occurrence of groundwater and development of solutional features have been significantly enhanced in the vicinity of crests and troughs of flexures (Kiersch, 1952; Kiersch and Hughes, 1952a,b; Henderson and others, 1959). The purpose of that study was to determine suitable sites along the Rio Grande for construction of a dam.

Cannon and others (1959) briefly noted Langtry Gypsum Cave and a fossil cave near the Pecos River Bridge on Route 90, 2.2 km northeast of the confluence of the Pecos River and the Rio Grande. They also referred to solutionally widened joints and faults in the vicinity of Langtry.

Freeman (1964a, 1968) described and mapped numerous areas of solution-subsidence in the southwest Edwards Plateau and commented on their origin. These

Table 9: Caves of the Langtry Area[a]

Cave	Length (m)	Depth (m)	Entrance Type	Stratigraphic Units	Termination
Langry Lead Cave	670	113	Doline	Boquillas, Buda, Devils River	Constriction
Emerald Sink	505	91	Pit	Buda, Devils River	Siphon
Langrty Quarry Cave	389	83	Doline	Boquillas, Buda, Devils River	Constriction
Langtry Gypsum Cave	180	20	Roadcut	Boquillas, Buda	Fill
Langtry East Gypsum Cave	260	21	Roadcut	Boquillas, Buda	Fill
Fisher's Fissure	198	81	Doline	Buda, Devils River	Siphon

[a]Based on descriptions and maps in Kunath and Smith (1968) and geologic map of Freeman (1964a).

zones of collapse range in size from a few meters to as much as 450 m wide and 5 km long. He related solution and subsidence to local stratigraphy and structure, and discussed the age of their development. The effect of zones of solution-subsidence on development of caves is discussed later in this chapter.

Reeves and Small (1973) briefly discussed karstic features such as caves, solutionally widened joints, and springs in relation to the groundwater of Val Verde County. The water-bearing properties of formations in this vicinity have been evaluated by Peckham (1963), Davis and Leggat (1965), Brown and others (1965), Reeves and Small (1973), and Rees and Buckner (1980). George and Frazier (1940b) inventoried wells and springs of the area.

Some of the results from this study which concern the Langtry area and are presented in this chapter have been previously published. Kastning (1980b, 1983a) contrasted the development of caves in the Langtry area with that of the Sonora area (Chapter 6), another deeply dissected section of the Edwards Plateau. He also investigated the influences of regional tectonism and fractures on the development of caves of the Langtry area and compared these findings with those from other areas of the Edwards Plateau (Kastning, 1981b, 1984) and from the Helderberg Plateau in New York and the Mississippian Plateau in Kentucky (Kastning, 1984).

DEVELOPMENT OF CAVES

Six caves were chosen for study (Table 9): Langtry Lead Cave, Emerald Sink, Langtry Quarry Cave, Langtry Gypsum Cave, Langtry East Gypsum Cave, and Fisher's Fissure. Four of these exceed 75 m in depth. Descriptive information, maps, and profiles are based on compilations by Reddell and Russell (1961b) and Kunath and Smith (1968).

Descriptions of Caves

Langtry Lead Cave

Langtry Lead Cave is 5.2 km north-northeast of Langtry and 1.2 km east of Eagle Nest Creek (Figure 18). It is denoted as "Big Tree Cave" on the Langtry 7.5-minute U.S.G.S. quadrangle. Total length and depth of the cave are 670 m and 113 m respectively (Figure 20).

The entrance of the cave is a vertically-walled doline, 14 m long, 8 m wide, and 12 m deep. From its bottom a 5 m-long crawlway leads to the top of a small room about 5 m high. From this room a passage leads southeast through several other rooms and beneath a series of domes up to 12 m in height. The passage drops in a stair-stepped fashion, and past a short side passage it enters the largest room, at the south end of the cave. This room is 30 m long, 18 m wide, and 6 to 12 m high. Near the entrance to the room another passage leads to the north. The third pit in the floor of this passage drops 5 m into a room 6 m in diameter and 6 m high. A 1 m-high and 46 m-long crawlway extends northwest from a hole in the floor of this room to a 6 m-deep drop that connects with a north-south trending passage. This passage ends 12 m to the south in a short pit and two small rooms, but it extends 15 m to the north as a stoopway connecting to the Hall of the Unicorns, a very straight passage trending 60 m to the north. The passage continues as a crawlway 18 m beyond this room but ends in clastic fill. On the west side of the Hall of the Unicorns, a side passage leads to the deepest drops in the cave which descend as a series of pits up to 30 m deep. The lowest point of the cave, 113 m below the entrance, is found at the south end of a passage that runs parallel to the Hall of the Unicorns.

The doline at the entrance is within the Boquillas Flags, but passages of the cave from the base of the doline to the large room at the south end of the cave lie within the Buda Formation. The remaining passages, including the crawlway leading from the large room, the Hall of the Unicorns, and the lowest levels are within the Devils River Formation (Figures 19 and 20).

During intense storms the cave receives runoff from the draw at the entrance. The crawlway between the large, southernmost room and the Hall of the Unicorns contains standing water during wet seasons. Bars of silt and gravel and drainage channels in the floors of some of the lower levels are evidence of occasional flooding of the cave.

Walls of upper-level passages are locally encrusted with gypsum crystals up to several centimeters in length. Gypsum flowers adorn a small room north of the large, southernmost room. Fossils of rudists protrude from the walls of the Hall of the Unicorns, giving it its name.

The topography in the vicinity of Langtry Lead Cave is moderately dissected. The draw at the entrance drops 33 m as it drains south to Eagle Nest Creek, a tributary to the Rio Grande.

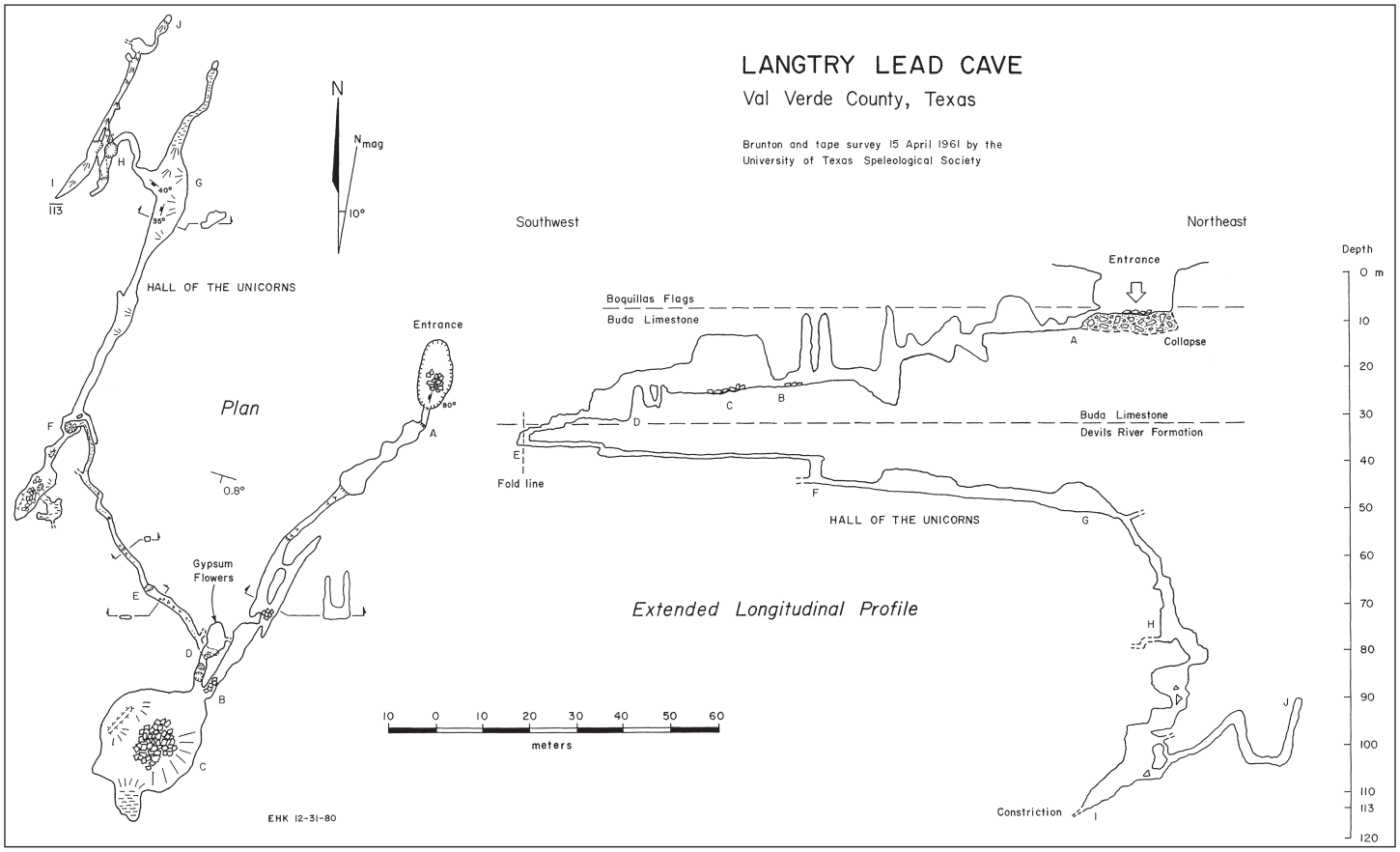


Figure 20: Map and profile of Langtry Lead Cave. Modified from Kunath and Smith (1968, p. 99, 101).

Emerald Sink

Emerald Sink is along Eagle Nest Creek, 1.05 km west-southwest of Langtry Lead Cave (Figure 18). It is denoted as "Emerald Cave" on the Langtry 7.5-minute U.S.G.S. quadrangle. The cave is 505 m long and 91 m deep (Figure 21).

The entrance is a 1.5 to 3 m-deep, 1 m-wide slot leading to a circular pit 3 m in diameter and 6 m deep. A passage trends in two directions from the bottom of the pit. To the southwest it is 3 to 6 m high, extending 58 m to a low room. A high fissure-like passage is encountered 6 m northeast of the entrance and extends north-northeast as a crawlway-stoopway for 107 m to a 37 m-deep pit.

Another fissure-like passage extends to the northeast from the entrance area for approximately 75 m to a 43 m-deep pit. A muddy crawlway, 0.3 to 1 m high, continues for 70 m as a series of segments of rectilinear passages to the top of a 15 m-deep pit that drops to a series of short, descending passages which end at a 1 to 2 m-deep pool of water. The pool, at 91 m below the surface, is the deepest point in the cave. It is nearly at the same depth as the bottom of the 37 m pit at the north end of the cave.

A 91 m-long linear passage, known as Backreef Crawl, trends south-southwest from a point 20 m northeast of the entrance. It is 0.3 m high for 6 m and then becomes a 6 m-high fissure-like passage that gradually becomes too narrow to pass.

The entrance of Emerald Sink is in the Buda Formation just below its contact with the overlying Boquillas Flags. The upper levels of the cave lie within the Buda Formation, near its lower contact with the Devils River Formation. The 37 and 43 m-deep pits and lower passages are within the Devils River Formation (Figures 19 and 21).

Sand and gravel deposits occur in the upper-level passages, and mud and silt in the lower passages. Locally, floors of passages are on bedrock and are devoid of sediment. Some large but inactive deposits of flowstone are present along the passage leading southwest from the entrance.

It is unlikely that the cave floods at the present time. Its entrance is 6 m above the level of Eagle Nest Creek, which lies about 100 m to the east.

Langtry Quarry Cave

Langtry Quarry Cave is 3.0 km northwest of Langtry and 150 m north of the road between Langtry and G 4 Canyon (Figure 18). It is denoted as "Quarry Cave" on the Langtry 7.5-minute U.S.G.S. quadrangle. The cave is 389 m long and 83 m deep (Figure 22).

The entrance to the cave is at the bottom of a 9 m-long, 6 m-wide, and 12 m-deep doline having vertical walls on three sides. The passage at the entrance consists of a series of parallel segments connected by short drops. At a depth of 33 m a low and wide passage extends southeast for 30 m and then northeast for approximately 120 m to a termination in breakdown and finer debris. This passage averages 6 to 12 m in width and 0.3 to 2 m in height.

Near the beginning of this main level a 25 m pit, named the Reef Pit, drops down to an intermediate level 57 m below the surface. The passage continues downward at a steep slope to 77 m below the surface where another level is maintained as a 60 m-long crawlway that ends in a 5 m-deep pit to the lowest level at 83 m below the surface. Several fissures cross the lower passage but end abruptly.

The doline at the entrance and the first 15 m of passage lie within the Boquillas Flags. Most of the passages near the entrance pass through the Buda Formation. The main level, containing most of the passages, has developed in the lowermost Buda Formation. The Reef Pit and the lower levels are within the Devils River Formation (Figure 22).

The main level contains a thick (up to 5 m) sequence of silt and clay that nearly fills the passage in many places. Some encrusted gypsum occurs on the walls and ceiling of this section. Fossils of rudists protrude several centimeters from the walls of the Reef Pit.

The cave is generally very dry; however, small pools of water have been observed on ledges of the Reef Pit. Langtry Quarry Cave is 9 m above the bottom of a small creek that is a normally dry tributary to Langtry Creek.

Langtry Gypsum and Langtry East Gypsum Caves

Langtry Gypsum and Langtry East Gypsum Caves are in a roadcut along Route 90, 1.6 km northwest of Langtry (Figure 18). These caves are not marked on the Langtry 7.5-minute U.S.G.S. quadrangle. Langtry Gypsum Cave is approximately 180 m long and 20 m deep and Langtry East Gypsum Cave is 260 m long and 21 m deep (Figure 23).

Entrances to both caves were artificially created during construction of a

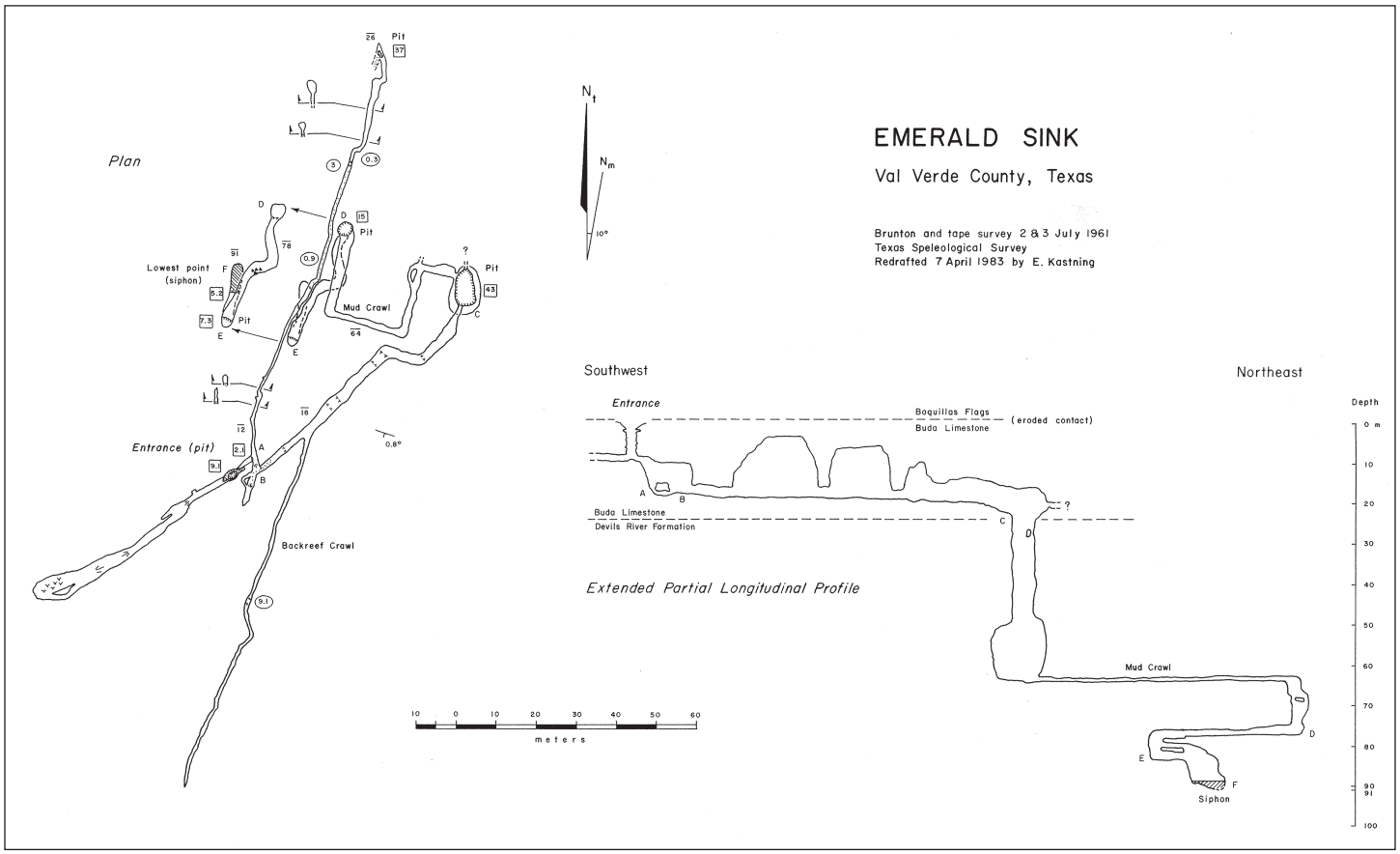


Figure 21: Map and profile of Emerald Sink. Modified from Kunath and Smith (1968, p. 81, 83).

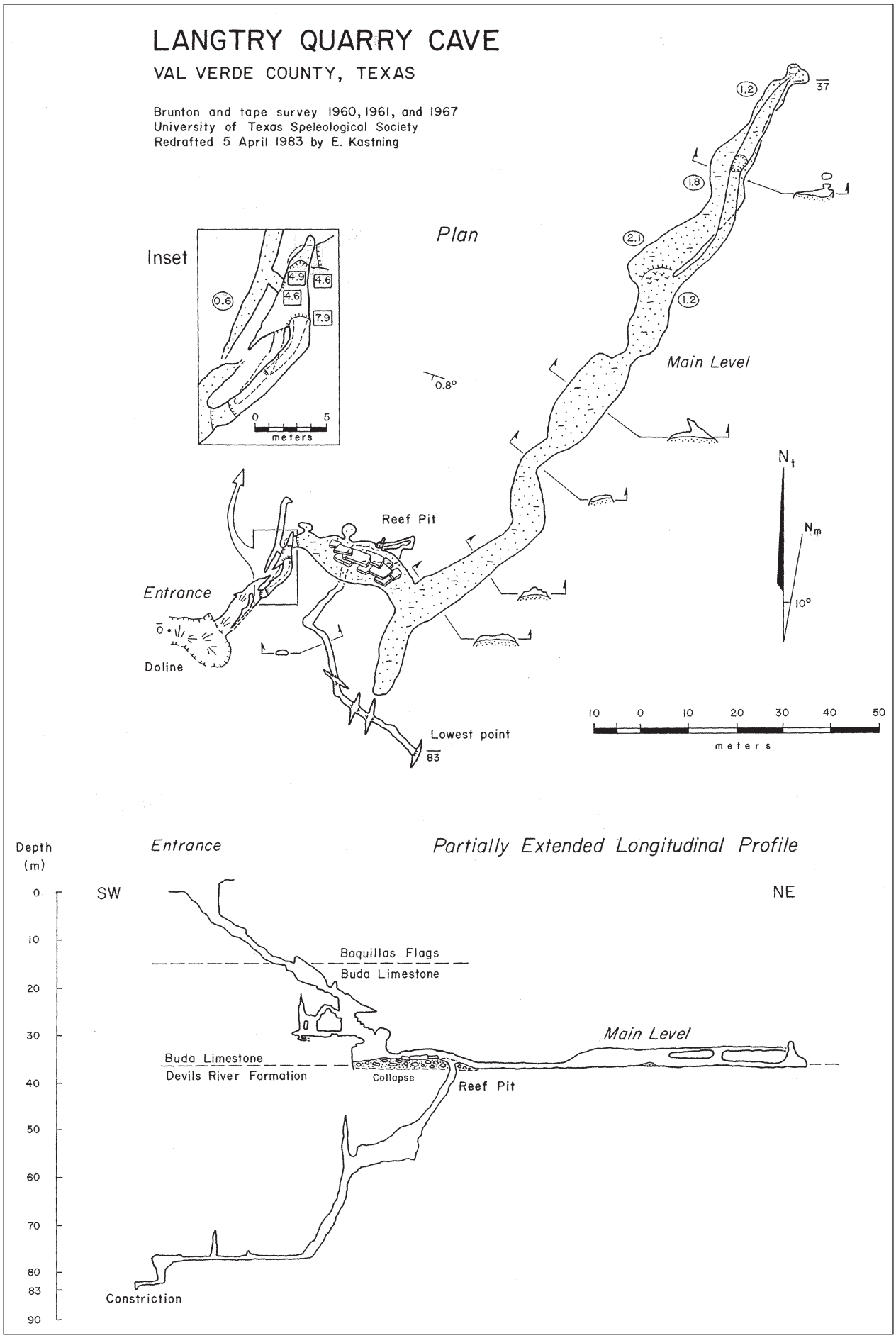


Figure 22: Map and profile of Langtry Quarry Cave. Modified from Kunath and Smith (1968, p. 105, 107).

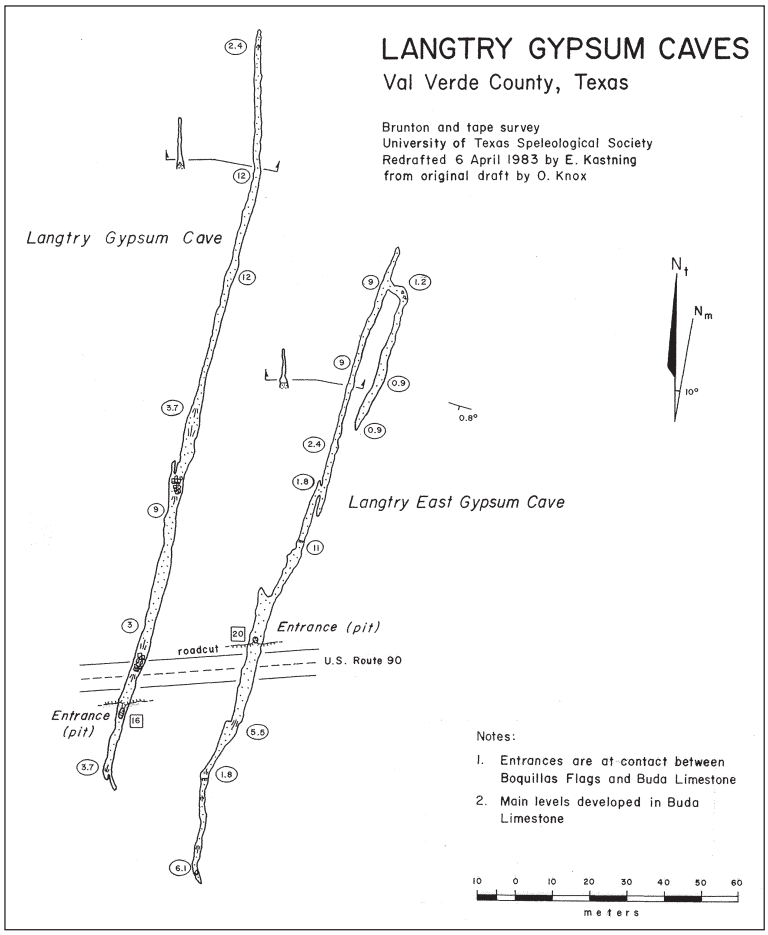


Figure 23: Maps of Langtry Gypsum Cave and Langtry East Gypsum Cave. Modified from Kunath and Smith (1968, p. 95).

highway. The entrance to Langtry Gypsum Cave (the western cave) is on the south side of the roadcut and that to Langtry East Gypsum Cave (the eastern cave) is on the north side. The entrances lead to drops of 16 m and 20 m respectively in each.

The caves are remarkably linear and parallel in plan-view, trending nearly north. Both run beneath the highway. Passages in both caves are generally fissure-like with heights of ceilings between 3 m and 11 m. Some short crawlways and sloping floors exist along these passages. The ends of the cave become too narrow for exploration or else terminate in sediment. A side passage branches from the north end of Langtry East Gypsum Cave and runs parallel to the main passage.

The entrances to both caves are at the contact of the Boquillas Flags and Buda Formation. The upper 2 m of the entranceway to Langtry Gypsum Cave is in the Boquillas, and the lower 1 m is in the Buda (see photograph on page 38 of Cannon and others, 1959). The main levels of both caves are within the Buda Formation (Figure 19).

Both caves are dry. Walls on the main levels are generally encrusted with white gypsum, ranging from delicate gypsum flowers, flakes, and hair-like clumps of gypsum whiskers to selenite needles and crystals. Some walls are almost entirely covered with slabs of gypsum, and mineral collectors have removed many of these deposits from Langtry Gypsum Cave. Slabs of breakdown are found locally on the floors of these caves.

The immediate area around the cave is moderately dissected. The surface is drained by an unnamed draw that flows to Langtry Creek, 0.8 km to the southeast and 75 m topographically lower than the entrances to the caves.

Fisher's Fissure

Fisher's Fissure is 3.8 km west-northwest of Langtry and approximately 250 m north of Route 90 (Figure 18). The cave is near the top of a bluff on the southeast side of the valley of Langtry Creek and is not marked on the Langtry 7.5-minute U.S.G.S. quadrangle. Fisher's Fissure is approximately 198 m long and 81 m deep (Figure 24).

The entrance is in an elongate doline, 12 m long and 4 m wide, from which two holes drop 5 to 6 m to a horizontal passage. This passage extends 70

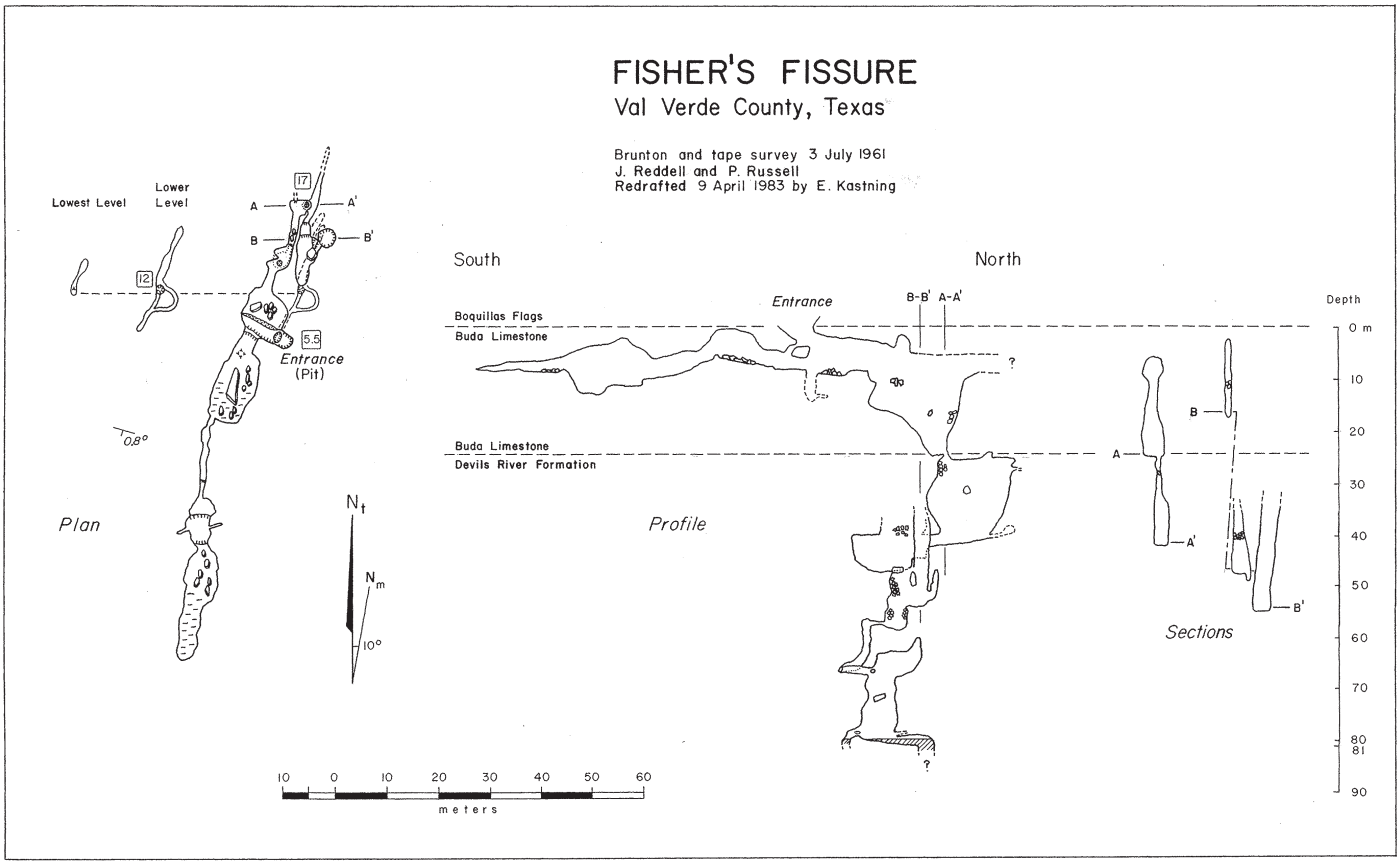


Figure 24: Map and profile of Fisher's Fissure. Modified from Kunath and Smith (1968, p. 87).

m to the south-southwest as a series of rooms up to 12 m long, 9 m wide, and 5 m high, and fissure-like passages 0.6 to 1 m wide and 6 m high. To the north-northeast the passage consists of a room 8 m in diameter leading to a series of fissures and pits descending to the lowest point in the cave (Figure 24).

The entrance to Fisher's Fissure is in the Buda Formation, just below its upper contact with the Boquillas Flags. The upper-level, horizontal passage lies near the middle of the Buda Formation and the series of fissures and pits at the northeast end of the cave descend 62 m into the upper Devils River Formation (Figure 24). A running stream was observed in the upper level southwest of the entrance in April 1968 (Kunath and Smith, 1968, p. 85); however, this level is usually dry. The cave is muddy in the lower levels and a pool of undetermined depth occurs at the deepest point. Slabs of breakdown litter the floors of the upper-level passage and breakdown is locally wedged between the walls of fissures and pits of the lower levels. The cave contains few speleothems.

Fisher's Fissure is in an area of moderate topographic dissection. Local baselevel is Langtry Creek, 75 m to the west and 13 m lower than the entrance to the cave.

Speleogenesis

Passages in the six caves described above have developed along distinct stratigraphic horizons as shown in the profiles (Figures 20, 21, 22, and 24) and in the geologic cross-section (Figure 19). Most segments of passages are linear in plan-view (Figure 20 through 24). This morphology suggests lithologic and structural control during development. The great depth of the caves suggests that they may have been developed in response to deep circulation of groundwater.

Lithologic Control

Each of the caves in the Langtry area has a distinct upper level developed in the Buda Limestone that is generally horizontal with little vertical relief. Some passages in the Buda are low and wide such as those in Emerald Sink, Langtry Quarry Cave, and Fisher's Fissure. Others are high and narrow (fissure-like) passages such as those in Langtry Lead Cave, part of Emerald Sink, Langtry Gypsum Cave, and Langtry East Gypsum Cave.

The overlying Boquillas Flags contain few passages. In those caves that have an entrance in the Boquillas Flags, development of passages has been through collapse of this thin-bedded incompetent unit into solutional cavities within the underlying Buda Limestone. The entrances to Langtry Lead and Langtry Quarry Caves are dolines resulting from such collapse. Here stoping of ceilings has progressed to the surface. At the entrance to Langtry Quarry Cave the zone of collapse has been guided by a northeast-trending fracture, and sagging beds overlying the passage at the entrance are visible in the wall of the doline (Figure 25; see also photograph in Fieseler and others, 1978, p. 105). Stoping of ceiling-beds in the Boquillas Flags at the entrance to Langtry Gypsum Cave is also evident (see photographi in Cannon and others, 1959, p. 38).

The lower Boquillas Flags are 90 percent dolomitic calcarenite and shale (Freeman, 1968, p. 11), and the high terrigenous, clastic content of this unit has reduced its solubility relative to that of the nearly pure limestone of the underlying Buda Limestone and Devils River Formation. During the establishment of flowpaths for groundwater and the onset of development of caves, the latter units were dissolved more easily than the Boquillas. Consequently, the Boquillas has acted as a resistant caprock overlying the cavernous limestones. Its resistance to solution has promoted formation of extensive, flat areas and cliffs on the surface, giving the area many rock-cut benches.

The horizontality of passages in the Buda Limestone suggests that groundwater may have been perched along stratigraphic horizons. Lithologic descriptions of the Buda and the underlying Devils River Formation indicate that argillaceous and dolomited bets occur locally along this boundary (Freeman, 1968). The Del Rio Clay is absent in the Langtry quadrangle because (1) it may have been removed by erosion following regional uplift along the Terrell Arch (northwest of Langtry) during the Late Cretaceous (Freeman, 1968, especially figure 3, p. 8), (2) it may never have been deposited here, or (3) it may be present as an equivalent facies of limestone in this area (K. Young, written communication). Diagenetic alteration of the upper Devils River Formation associated with subaerial exposure, and subsequent burial during the deposition of the Buda, may have resulted in a less soluble bed upon which the upper levels of the cave developed.

Nearly all vertical development of the caves studied here has occurred in the Devils River Formation. The uppermost beds of this unit are clayey, thin beds. However, beds below this consist of very thick, pure limestone. Massive beds, vertically uniform lithology, and high calcitic content of the Devils River Formation have promoted dissolution along vertically extensive fractures rather than along bedding-plane partings. As a result, passages in the Devils River Formation tend to be fissures and pits developed along fractures of short, horizontal extent. This is in contrast to low and wide passages along bedding-plane partings in the Buda Formation.

Zones of recrystallization, dolomotization, and chert occur locally in the Devils River Formation. Various levels in deeper parts of some of the caves may be partly the result of perching of groundwater on such zones. Most horizontal segments of passages in the Devils River Formation are short (less than 100 m long). Some lower levels may be related to lithologic variability within the reefal facies of the Devils River.

Structural Control

Relationship to Fractures. Development of caves in the Langtry area has been strongly guided by fractures. Reddell and Russell (1961b) made a preliminary analysis of trends of passages within caves and Smith (1971b) showed that these correlated with lineaments mapped by Freeman (1964a).

Rose diagrams showing trends of passages in each of the six caves under study are shown in parts A through E of Figure 26. The diagrams for Langtry Gypsum Cave and Langtry East Gypsum Cave have been combined (Figure 26D). A composite diagram, Figure 26F, shows the trend of passages for all six caves.

Lineaments mapped by Freeman (1964a) in the vicinity of Langtry are shown in Figure 27, and a rose diagram of these is presented in Figure 26G. In addition Leonard (1977a) mapped lineaments from controlled, aerial photomosaics at a scale of 1:20,000 (on file at the Bureau of Economic Geology, University of Texas

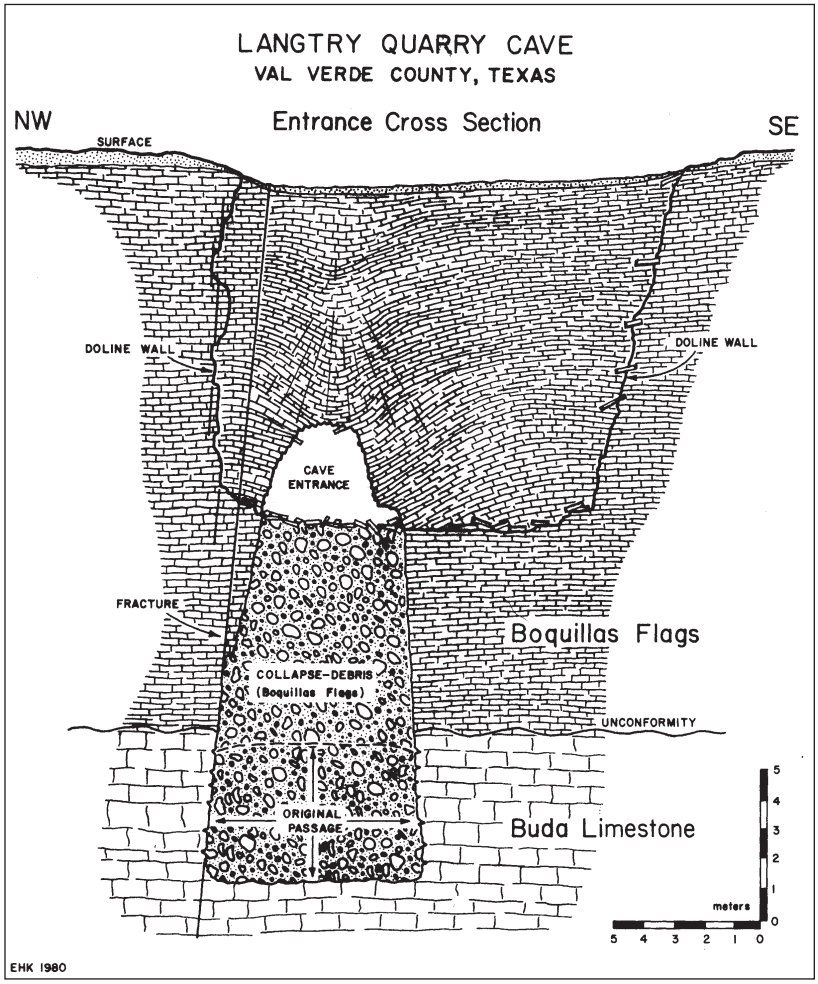


Figure 25: Cross-sectional sketch of entrance to Langry Quarry Cave, showing collapse of Boquillas Flags into underlying cavity inthe Buda Formation.

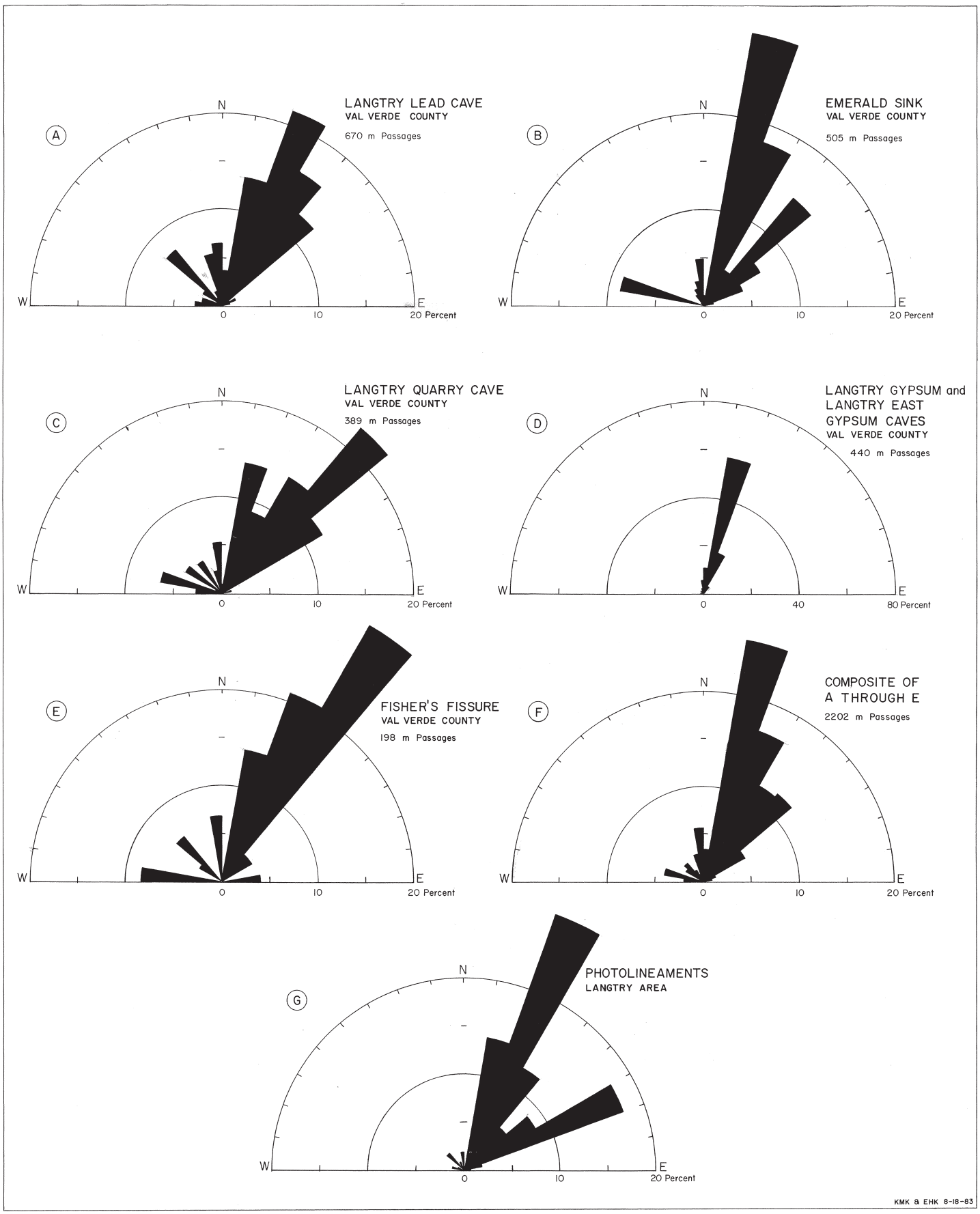


Figure 26: Rose diagrams of linear segments of passages in caves and mapped photolineaments, Langtry area. Linear segments of passages are shown in(A) Langtry Lead Cave, (B) Emerald Sink, (C) Langtry Quarry Cave, (D) Langtry Gypsum and Langtry East Gypsum Caves, and (E) Fisher's Fissure based on measurements from maps of Figures 20 through 24. (F) is a composite diagram showing linear segments of passages of all six caves. (G) represents photolineaments mapped by Freeman (1964a).

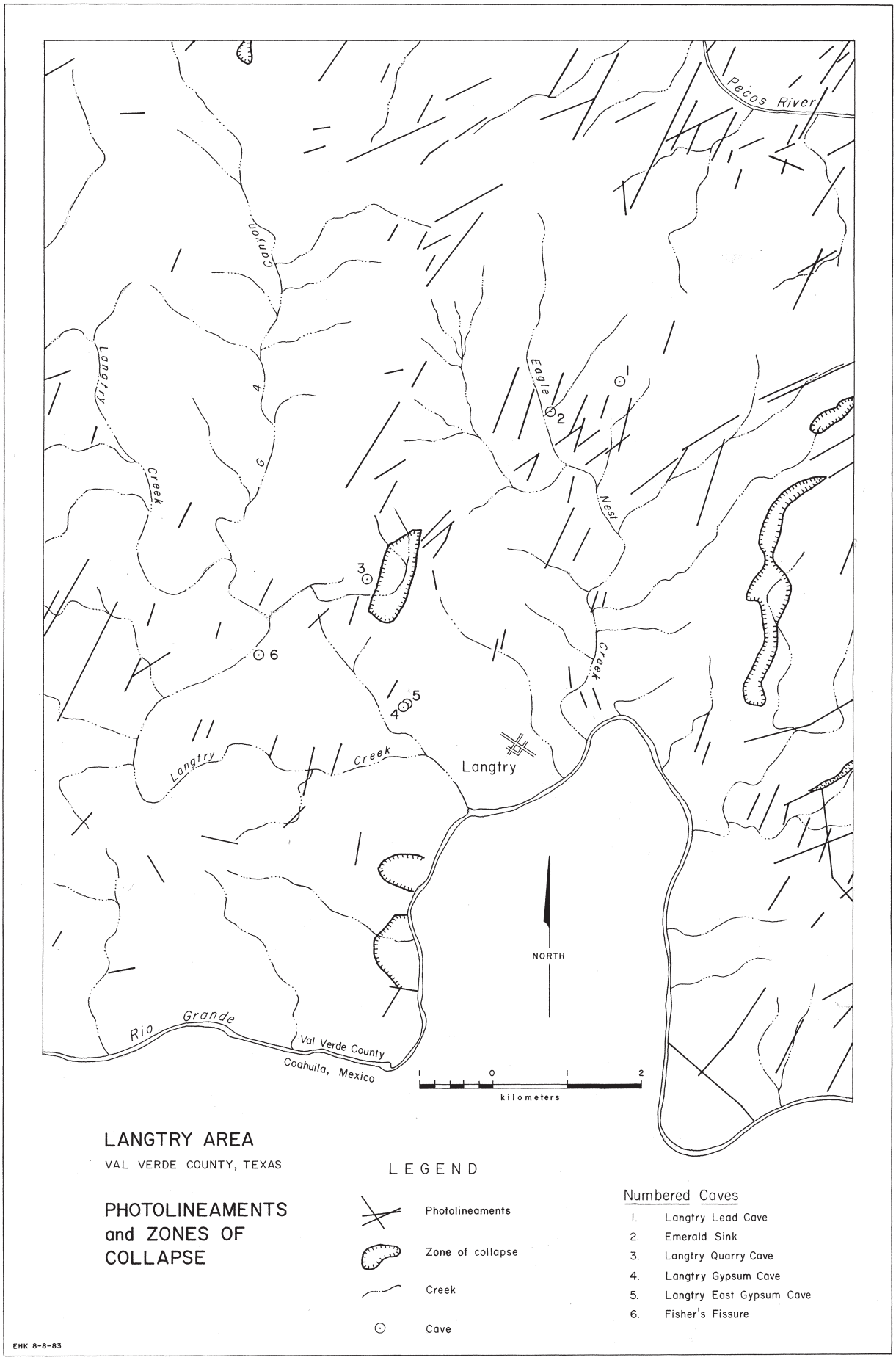


Figure 27: Map of lineaments and large-scale features of solution and subsidence in the Langtry area. Map area is the same as in Figure 18. Based on map by Freeman (1964a).

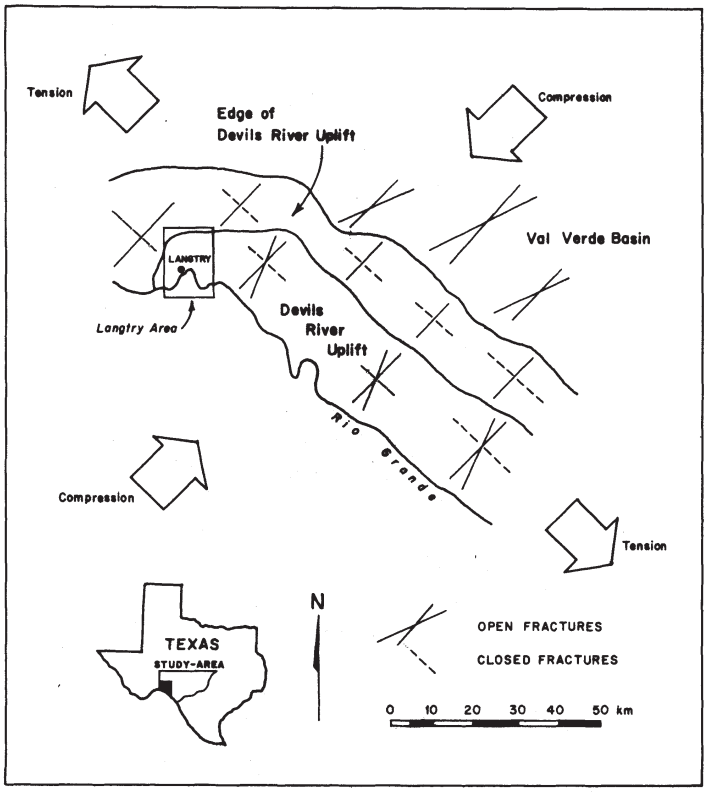


Figure 28: Compressional and tensional forces acting along the structural trend of the former Devils River Uplift during the uplift of the Edwards Plateau in the Late Cretaceous and Early Cenozoic. Open and closed trends of fractures resulting from this event are indicated. Based on Leonard (1977a, figures 13, p. 13 and 32, p. 65).

at Austin). He also measured joints in the field, and rose diagrams of these are available in his thesis (Leonard, 1977a).

Comparisons among rose diagrams of Figure 26 result in the following observations (letters A through G to individual rose diagrams of Figure 26):

1. Trends of cave passages (A through F) correlate very well with mapped photolineaments (G). They also agree well with trends of photolineaments (G). They also agree well with trends of photolineaments and joints mapped by Leonard (1977a).

2. Orientations of Cave passages (A through E) and photolineaments (G and those of Leonard [1977a]) have dominantly northeastern components. However, joints mapped on the ground by Leonard (1977a) have a strong northwest set as well as northeastern sets.

3. Orientation of cave passages (A through E) generally range from N 10°-60° E. Some caves have passages aligned strongly within the N 10°-20° E interval. (Emerald Sink [B}, Langtry Gypsum Cave [D], and Langtry East Gypsum Cave [D]), and others are aligned strongly within the N 30°-40° E or N 40°-50° E intervals (Langtry Quarry Cave [C] and Fisher's fissure [E]).

4. Orientations of photolineaments (G and those of Leonard [1977a]) generally vary between N 10° E and N 60° W.

As discussed previously, the two northeastern trends of lineaments may represent sets of conjugate fractures of small dihedral angle that result from simultaneous compression and extension acting at 90 degrees to each other (Figure 28). The northeast-trending fractures, which are nearly parallel to the compressional forces and perpendicular to the extensional forces, have undergone separation during tectonism.

The northwest-trending set of fractures is perpendicular to forces of compression (Figure 28) and may have resulted from slight folding along the northwest-southeast axis of the Devils River Uplift during the Late Cretaceous to Early Cenozoic (Fowler, 1956, 1957; Leonard, 1977a). If the northwest0trending fractures are compressional features, it is expected that they did not separate appreciably during tectonism.

Under the foregoing interpretation it follows that northeast-trending fractures would tend to be generally open fractures and the northwest ones closed. Therefore, infiltration and circulation of groundwater, and ensuing development of caves, would occur preferentially along the northeast-trending fractures. This thesis is borne out by examination of serial photographs (e.g. photograph in Cannon and others, 1959, p. 33). Northeast-trending lineaments are readily visible on the photographs because (1) they have been widened by solution in response to their openness, and (2) rubble, soil, and moisture, which have been trapped in widened joints, support a relatively

dense vegetation (compared to nearly terrain) that is visible as dark lineaments in the imagery. Conversely, northwest-trending lineaments have not been solutionally widened, do not support vegetation as well, and are difficult to recognize in the imagery. These observations are consistent with those of Lattman and Olive (1955), who studies patterns of fractures in the carbonate terrain of the western Stockton Plateau.

Where both compressionally and tensionally induced fractures occur, measuring joints on the ground gives a much more representative sample of all orientations. Leonard (1977a, p. 20) has identified a major northwest (compressional) component throughout southern Val Verde County in the vicinity of the Devils River Uplift. Northwest-trending joints have a smaller deviation in orientation than do the northeast-trending joints. Consequently, some orientations of joints of the northwestern set occur at greater frequencies than do those of the northeastern sets. Joints of the northwestern set, which parallel the fold-axis of the Devils River Uplift, are relatively long compared to those of the northeastern sets, and this may also increase a frequency distribution based on length.

Rose diagrams (Figure 26) indicate that passages of caves generally follow fractures of either of the two conjugate sets (N 10°-30° E or N 40°-50° E). A comparison of rose diagrams of individual caves (Figures 26, A through E) and lineaments shown to be near the caves in Figure 27 confirm this relationship. Very little length of passages has developed along northwest-trending joints (Figure 26F).

Fractures of any given trend are not confined to particular stratigraphic horizons. A glance at maps and profiles of the deep caves in the Langtry area (Figures 20, 21, 22 and 24) shows that passages have developed along fractures of the same orientation throughout the cavernous, vertical sequence, including the Boquillas Flags, and Buda and Devils River Formations, suggesting that all units in the sequence have responded similarly to tectonic stresses. This is particularly true of the strata of the Devils River, where homogeneity in lithology and competency throughout the thickness has promoted uniform fracturing throughout the formation. In Emerald Sink and Fisher's Fissure some passages in the Buda and Devils River Formation overlie one another along the same fracture. Vertical continuity of fractures has enhanced deep circulation of groundwater and concomitant development of caves.

The density of fractures varies over the Langtry area (Leonard, 1977a, figure 16, p. 38; Freeman, 1964a). Fractures are more frequent along the strikes of flexures, where structural contours are closely spaced (Freeman, 1964a). An increase in density of fractures would be expected to enhance development of caves provided other factors do not change. Indeed, the two longest caves in this vicinity, Langtry Lead Cave and Emerald Sink, are located in an area of high density of fractures (Figure 27). All six caves are located within 2 km of flexural axes (Figure 19). It is interesting to note that Kiersch and Hughes (1952a,b), in a study of the Rio Grande area, found that solutional features, mainly dolines, and the occurrence of groundwater are associated with crests and troughs of flexures, in zones where the density of fractures owing to tectonism is greatest.

Modification by Collapse. Two episodes of subsidence and collapse have been recognized in the Langtry area (Freeman, 1968, p. 22-24). Large-scale subsidence (up to 1200 m in horizontal extent) occurred between the Late Cretaceous and early Pleistocene. These features are younger than the rocks mapped as Austin Chalk by Freeman (1964a) and Barnes (1977), as that formation has been affected by subsidence. On the other hand, this episode probably predated the incision of the Rio Grande and Pecos River because the piezometric surface must have been near the surface to allow large-scale, phreatic solution to occur high in the stratigraphic sequence. A second period of subsidence and collapse has been occurring from the Pleistocene to the present and is related to structural collapse of caves caused by recent erosion of the surface.

Of the six caves investigated, only Langtry Quarry Cave is near a large zone of subsidence as mapped by Freeman (1964a) (Figure 27). Here, collapse of incompetent beds of the Boquillas Flags (Figure 25) may be related to structural weaknesses resulting from previous, large-scale subsidence. However, a more plausible mechanism is that overlying beds of Boquillas Flags collapsed into a large cavernous opening at or near the top of the Buda Formation (Figures 22 and 25). Collapse progressed upward through stoping until a mechanically stable configuration was achieved. The entrance to Langtry Lead Cave (Figure 20) may have formed in a similar fashion. The process and mechanics of collapse of beds into underlying solutional cavities is described in more detail in Chapter 10 for the case of Natural Bridge Caverns in Comal County.

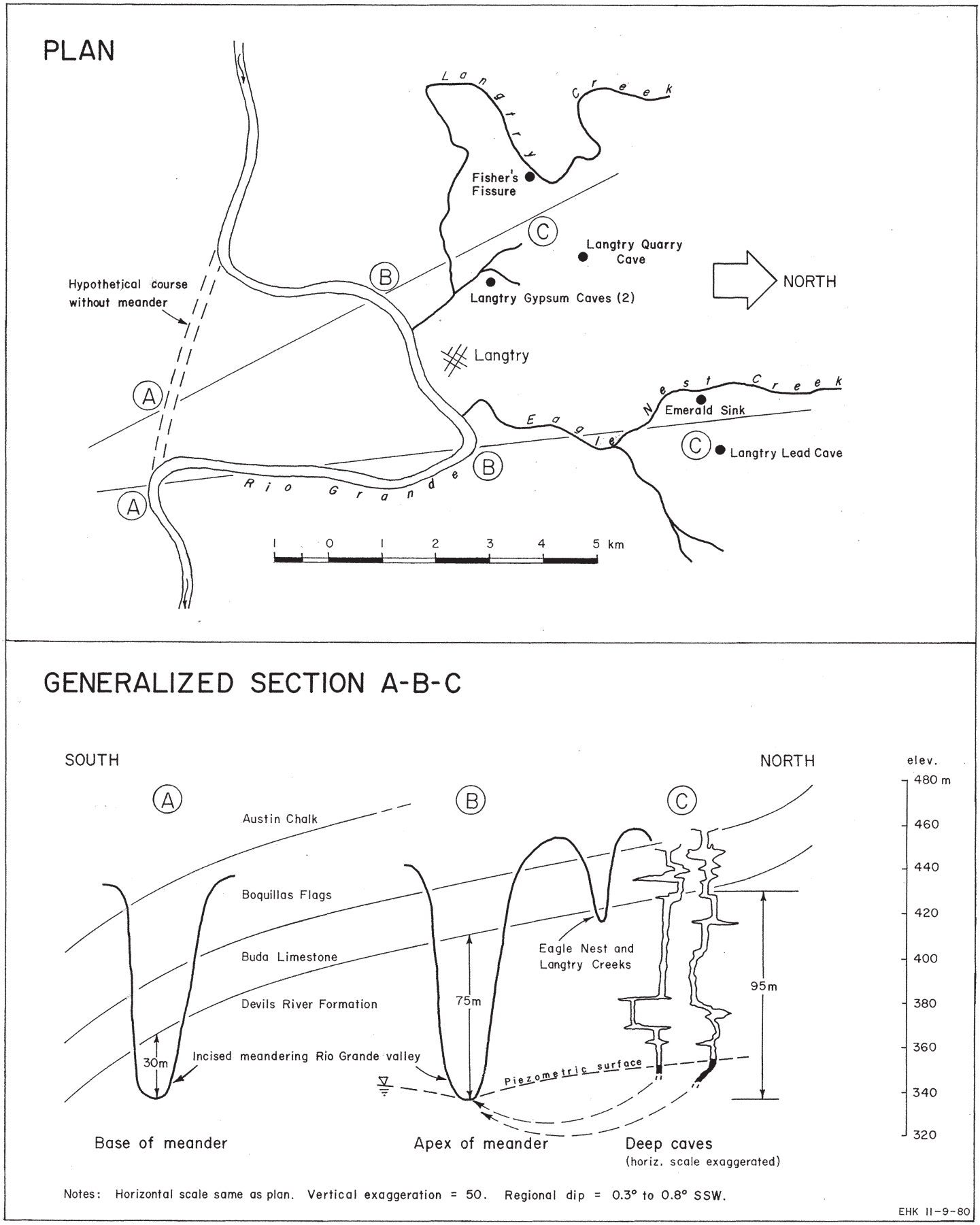


Figure 29: Development of caves as related to local baselevels and to the incision of the Rio Grande in the vicinity of Langtry. (A) Along the main course of the Rio Grande incision has penetrated only 30 m of the Devils River Formation. (B) At the north end of the large meander near Langtry the river has incised 75 m into the Devils River Formation. (C) The cavernous zone of the Devils River Formation reaches a potential thickness of over 95 m in the vicinity of Langtry Lead Cave and Emerald Sink. This generalized section is based on stratigraphic and structural information on the map by Freeman (1964a), the geologic section of Figure 19, and maps and profiles of caves (Figures 20 through 24).

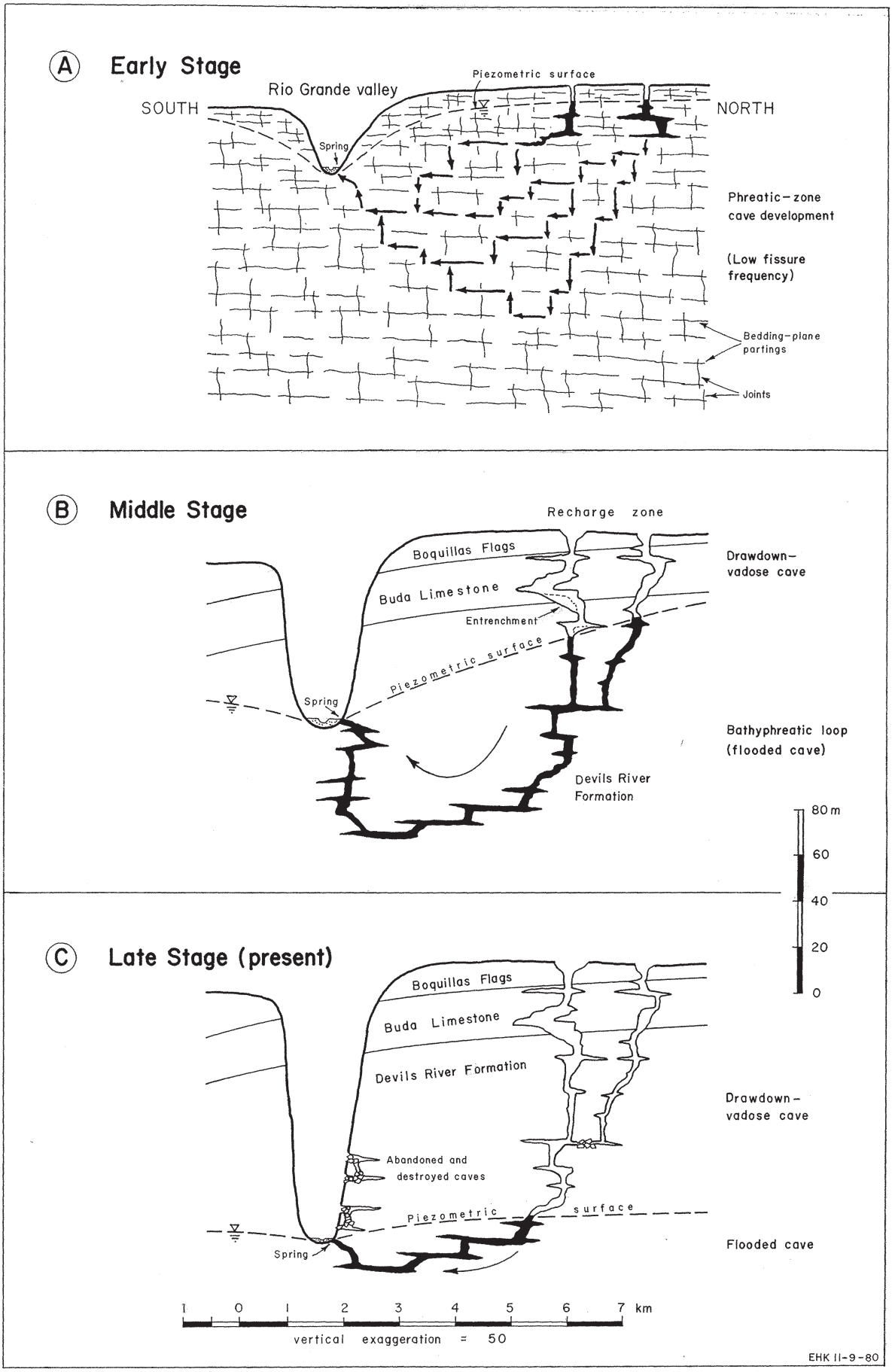


Figure 30: Schematic profile of bathyphreatic development of caves in the Langtry area: (A) deep phreatic circulation and incipient development of caves; (B) drop in piezometric surface accompanying incision of the river and vadose modification of abandoned, upper levels; and (C) drawdown-vadose cave in relation to present-day baselevel. Based in part on Ford and Ewers (1978, figure 5, p. 1789).

Hydrogeologic and Topographic Controls

That caves of the Langtry area are the deepest in the area of study suggests that circulation of groundwater was deep during their development, and this in turn implies that sufficient topographic relief must have existed during speleogenesis to provide impetus for circulation. Topographic relief in the vicinity of the caves resulted from the incision of the Rio Grande and its tributaries, Langtry and Eagle Nest Creeks.

It is reasonable to question why most of the deepest known caves in the Edwards Plateau are clustered in a small 5 km by 5 km area just north of Langtry. Similar lithic and structural conditions exist elsewhere in Val Verde County, and topographic relief is sufficient for deep caves along other reaches of the Rio Grande. The key to this problem is the large meander in the Rio Grande that extends 5.5 km north from the relatively straight course that the river maintains upstream and downstream from the meander (Figure 18).

The relationships among the circulation of groundwater in the vicinity of the Rio Grande, the caves, and the meander are easily visualized by referring to the schematic cross-section of Figure 29. The elevation of the Rio Grande (prior to impoundment by the Amistad Dam at Del Rio), as it crosses the Langtry 7.5-minute quadrangle, drops from 346 to 340 m. Along the meander the Rio Grande drops less than 2 m. The surrounding uplands are of relatively uniform elevation. Therefore, there has been very little variation in depth of topographic incision by the river along the meander.

However, the *stratigraphic* incision of the Rio Grande varies markedly along the meander. The river has breached the Boquillas Flags and Buda Formation and penetrated up to 75 m into the underlying Devils River Formation at the north end of the meander (point B in Figure 29). However, because the regional dip is to the south-southwest, the Rio Grande has penetrated only 25 to 30 m of the Devils River Formation along the southern reaches of the meander (point A in Figure 29). Therefore an additional 45 to 50 m of the Devils River sequence, above the level of the Rio Grande, was made available for the circulation of groundwater and speleogenesis in the vicinity of Langtry. Assuming that the groundwater responsible for the development of the northernmost caves, Langtry Lead Cave and Emerald Sink, flowed to the Rio Grande,the effective speleogenetic thickness of the Devils River was at 95 m (point C in Figure 29). Therefore, deep incision of the river into the Devils River Formation in the vicinity of Langtry provided the primary impetus for deep circulation of water and development of caves.

Bathyphreatic Development of Caves

The caves of the Langtry area are excellent examples of the bathyphreatic state (State 1) of the "Four State Model" of phreatic and water-table systems of caves recently proposed by Ford and Ewers (1978). The example of a single bathyphreatic loop has been discussed several times by Ford (1971a,b,c, 1977a,b) and encompasses the theoretical considerations of Davis (1930) and Bretz (1942) which describe development in the deep-phreatic zone.

Ford and Ewers (1978, p. 1789-1790) indicate that a bathyphreatic cave may be composed of several segments, along bedding-plane partings and joints, that together constitute a single conduit, or loop, from an area of recharge to a point of discharge (spring). Single, deep, phreatic loops result within host rocks of minimum "fissure frequency" (i.e. frequency of all kinds of permeable partings, whether they are bedding-plane partings or fractures). In the Langtry example, the massive-bedded Devils River Formation had relatively few open bedding-plane partings that allowed the flow of groundwater. Vertically oriented fractures were prevalent in the area, but most were spaced tens of meters apart.

A schematic diagram of a typical bathyphreatic cave in the Langtry area, based on the conceptual model of Ford and Ewers (1978), is shown in Figure 30 with scales appropriate to the area. Initially, prior to incision of the Rio Grande, the entire cave developed within the phreatic zone. However, as the piezometric surface dropped, the upper levels of the cave were drained of water and became passages of what Ford and Ewers (1978) term "drawdown-vadose caves."

Deep-phreatic conditions are corroborated by water-wells. Henderson and others (1959, p. 176) state that some deep wells driven into the Cretaceous rocks north of the Rio Grande, along the reach between Indian Creek and Del Rio, have artesian rises from 60 to 90 m.

Chronology of Speleogenesis

The chronology of the development of caves in the Langtry area is summarized as follows, based on the foregoing observations and interpretations:

1. During the Early Cretaceous the Devils River Formation was deposited in Val Verde County along the structural trend of the Devils River Uplift.

2. Regional uplift during the early Late Cretaceous resulted in nondeposition of the Del Rio Clay in the vicinity of Langtry and perhaps erosion of the topmost beds of the Devils River Formation. Some karstification and development of non-integrated solutional cavities may have occurred in the Devils River Formation at this time.

3. The Buda Formation and Boquillas Flags (including the equivalents of the Austin Chalk) were deposited during the remainder of the Cretaceous as further uplift occurred in the vicinity of the Devils River structural trend. As a result, these units thickened slightly to the south in the Langtry area (Freeman, 1968, figures 4 and 5, p. 10, 13).

4. Uplift of the Edwards Plateau, beginning at the close of the Cretaceous, created three sets of fractures, one compressional and two tensional. These enhanced the circulation of groundwater and promoted the dissolution of carbonate rocks. Large areas subsided concomitantly with solution, beginning at this time and continuing into the Tertiary.

5. Continued uplift during the Tertiary promoted erosion and stripping of Upper Cretaceous rocks from the area. Tensional (northeast-trending) fractures became widened by dissolution.

6. Incision of the Rio Grande and Pecos River began in the middle to late Tertiary. Meanders of rivers were preserved as subsequent features during incision.

7. Incision continued during the Pleistocene. However, there may have been pauses in downcutting of the Rio Grande and Pecos River, as suggested by deposits from former fluvial terraces at elevations higher than present-day levels of terraces (Freeman, 1964a; Leonard and Frye, 1962). Circulation of groundwater was integrated into phreatic loops along bedding-plane partings and fractures (Figure 30A). Upper-level passages of caves in the Buda Limestone probably developed in the shallow-phreatic zone during the early Pleistocene.

8. As river incision progressed to its present depth, bathyphreatic loops developed in the vicinity of the large meander in the Rio Grande near Langtry, where incision cut well into the soluble Devils River Formation (Figure 30B). Under steep hydraulic gradients, bedding-plane partings and fractures of wide spacing became segments of deep loops extending from points of recharge to low-elevation springs along the Rio Grande and its tributaries.

9. The piezometric surface dropped in response to lowered base levels. Drawdown-vadose caves were established in the upper levels of the deep caves (Figure 30C). Floors of passages became entrenched by vadose flow in upper levels.

10. Collapse of the thin-bedded Boquillas Flags began where passages in the Buda Limestone were close to the surface, and where the Boquillas Flags were eroded to thin exposures above the caves.

SUMMARY

Caves developed in the Devils River Formation in the Langtry area are anomalously deep. The Devils River Formation is a stratigraphically homogeneous and highly pure limestone in which fractures are few, widely spaced, and vertically extensive. Incision by a meander of the Rio Grande and its tributaries, extending well updip into the stratigraphic section, has locally exposed up to 75 m of the formation. This in turn has steepened local, hydraulic gradients and promoted bathyphreatic circulation and development of deep caves along the fractures. Horizontal passages are relatively short and are connected by vertical shafts and steeply inclined corridors. The latter exhibit entrenched floors which are characteristic of subsequent enlargement under vadose conditions.

6
SONORA AREA

DESCRIPTION OF AREA

The Sonora area lies wholly within and includes a large part of the western half of Sutton County (Plate 1). The area is bounded on the south and north along lines of 30°22'30" and 30°37'30" north latitude respectively, and on the east and west along lines of 100°30'00" and 100°52'30" west longitude respectively. These boundaries coincide with the outer borders of the Caverns of Sonora, Sonora, Sonora SE, Hudspeth Draw, Dunbar Draw NW, and Dunbar Draw NE 7.5-minute U.S.G.S. quadrangles (Figure 31). The area is drained by Granger Draw, Dry Devils River, Cusenbary Draw, and Dunbar Draw, which are upstream tributaries of Devils River within the Rio Grande Basin (Plate 2).

The most detailed, published, geologic map of the area is at a scale of 1:250,000 (Barnes, 1981b). Other maps of this vicinity include those of Leggat (1954a,b) and Brown and others (1965).

Exposed rocks in the Sonora area consist largely of beds of the Segovia Formation of the Edwards Group. Locally, exposures of Buda Limestone cap some high hills, and deposits of Quaternary alluvium cover floors of major fluvial valleys.

The topography of the Sonora area is generally developed on the upper and middle parts of the Segovia Formation, which consist of sequences of dolomitic limestone and marl. The uppermost beds are medium- to course-grained, cherty, medium-bedded to massive, and contain abundant mollusc fragments, spar, and scattered rudistid corals. There is also a fair amount of mudstone. In this region one or more thick beds of limestone form prominent, resistant ledges that provide a caprock over underlying beds of the Segovia. The middle beds are dolomitized micrite and marl, are thin- to thick-bedded, soft, and porous, and tend to weather recessively in contrast to upper Segovia beds. A deeply weathered and altered, fine to coarse, recrystallized dolomitic limestone, lying between the *Gryphaea* Bed and Orr Ranch Bed (Rose, 1972), weathers to a distinctive, open, grassy slope separating the caprock-forming, upper beds from the recessive, middle beds (Curry, 1934; Rose, 1968, 1972). Differences in topography, soils, and vegetation among these units facilitates their recognition from aerial photographs. I have mapped the geology in the vicinity of three major caves (Figure 31) using imagery published in Wiedenfeld and McAndrew (1968).

The Segovia Formation in the Sonora area lies generally parallel to the pre-Cretaceous surface (Cartwright, 1932, figures 1 and 3; Holmquest, 1955; Rall and Rall, 1958a; Brown and others, 1965, plate M10; Blank and others, 1966, figure 5; Walker, 1979, figures 9 and 13). The beds are nearly horizontal, striking approximately N 45° W and dipping only 0.1° to 0.2° to the southwest.

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

A descriptive catalog of caves of Sutton County has not yet been compiled; however, substantial information is on file with the Texas Speleological Survey. Over 50 known caves are distributed throughout the county, but most are less than 100 m long. Of six caves in the county known to exceed 300 m in length, three (Caverns of Sonora, Felton Cave, and Silky Cave) lie within the designated Sonora area and are the basis of speleogenetic interpretation for this locale.

Speleological activity in the Sonora area was minimal until 1955, when explorers discovered the well-decorated, inner passages of Mayfield Cave (Crisman, 1956; Tandy, 1962). The cave was commercially developed and opened to the public in 1960 as the Caverns of Sonora (Papadakis, 1977). In the meantime cavers began mapping and exploring Felton Cave (anonymous, 1959, 1960; Simpson, 1959; Russell, 1962). Profuse helictites, coralloids, and other speleothems of the Caverns of Sonora (Figure 32) caught the attention of several investigators, who have described them and have speculated on their origin (Semken, 1957; Halliday, 1959; Quinlan, 1959; Burch, 1967; Moore and Bukry, 1968). Frank (1963, 1964, 1965a, 1971) and Lundelius (1967) studied sediments and vertebrate remains from the entrance area of Felton Cave and found that the depositional history of detrital material in the cave is representative of entrance-facies of other central Texas caves. These sediments typically include (1) clay and quartz from surficial soil, (2) lithic fragments of limestone and calcite derived from the bedrock and broken speleothems, and (3) fragments of bone deposited by predators which occupied the cave.

The United States Department of Agriculture, Agriculture Research Service, undertook a reconnaissance study of the geology and groundwater of the Lowrey Draw watershed just northeast of Sonora (Blank, 1963; Knisel, 1965; Blank and others, 1966). This study indicated that large volumes of water are rapidly transmitted underground through solutional openings in the bottoms of fluvial channels and flood-retention reservoirs. Recharge from nondissected uplands descends through dolines and joints to permeable, dolomitic beds underlying lowland areas. There is little primary permeability in the limestone and dolostone of this region, and secondary permeability is due to joints, caves, and "honeycomb" rocks characteristic of weathered rudistid reefs. Blank and others (1966) indicate that secondary (solutional) openings are numerous near the surface and decrease significantly in number and size at depths greater than 30 m. Levels of water in wells and the position of the regional, piezometric surface are documented in Blank and others (1966), Muller and Pool (1972), and Walker (1979).

Some of the results of investigations described in this chapter have appeared in published papers. The origin of long, horizontal caves in the deeply dissected terrain of the Sonora area, which is typical of much of the western Edwards Plateau, has been contrasted to development of caves in the Langtry area (Chapter 5), where selected stratigraphic incision of a major river has promoted deep caves (Kastning, 1980b, 1983a). Moreover, the structural implications of speleogenesis in the Sonora area has been compared with that of three other areas of the Edwards Plateau and with that of the Helderberg Plateau in New York and the Mississippian Plateau in Kentucky (Kastning, 1981b, 1984)

DEVELOPMENT OF CAVES

Descriptions of Caves

Caverns of Sonora

The Caverns of Sonora are 16.5 km west of Sonora and 2.3 km southeast of route 1989 and are indicated on the Caverns of Sonora 7.5-minute U.S.G.S. quadrangle (Figures 31 and 33). The surveyed length of the cave is currently 2.2 km making it one of the longest caves in Texas (Figure 34). The lowest point in the cave is approximately 45 m below the natural entrance. End-to-end extent of the cave is 656 m.

The natural entrance to the cave was originally a small hole in bedrock on the side of a valley. It was too small to enter and was later chipped to a larger size to permit exploration and enlarged to its present size upon commercial development. An artificial entrance, located 385 m to the northeast, serves as an exit for tours.

The general trend of passages in the cave is N 32° E, from the natural entrance to the northernmost sections (Figures 33 and 34). One section, known as the Diamond Room Passage, trends S 22° W from where it branches from the rest of the cave.

An 18 m-long passage descends from the natural entrance and intersects a 245 m-long passage, known as the Ledge Route, extending from a point 43 m northeast of the entrance to the top of an 18 m-deep pit (Pit Room). At the Register Room, 30 m from the entrance passage, a short perpendicular corridor descends and connects with a passage parallel to the Ledge Route but at a lower level. After approximately 100 m, this route intersects the Auditorium, a passage whose 8 m-wide by 9 m-high cross-sectional dimensions are among the largest in the cave (Tandy, 1960, 1962, figure 2, p. 34). This passage terminates in the Pit Room at floor level. The passages described above, totaling 600 m in length, represent all that was known of Mayfield Cave until late in 1955 when the Ledge (Devil's Delight) was first crossed, and the continuation of the upper-level Ledge Route was explored. A photograph of the Ledge is in

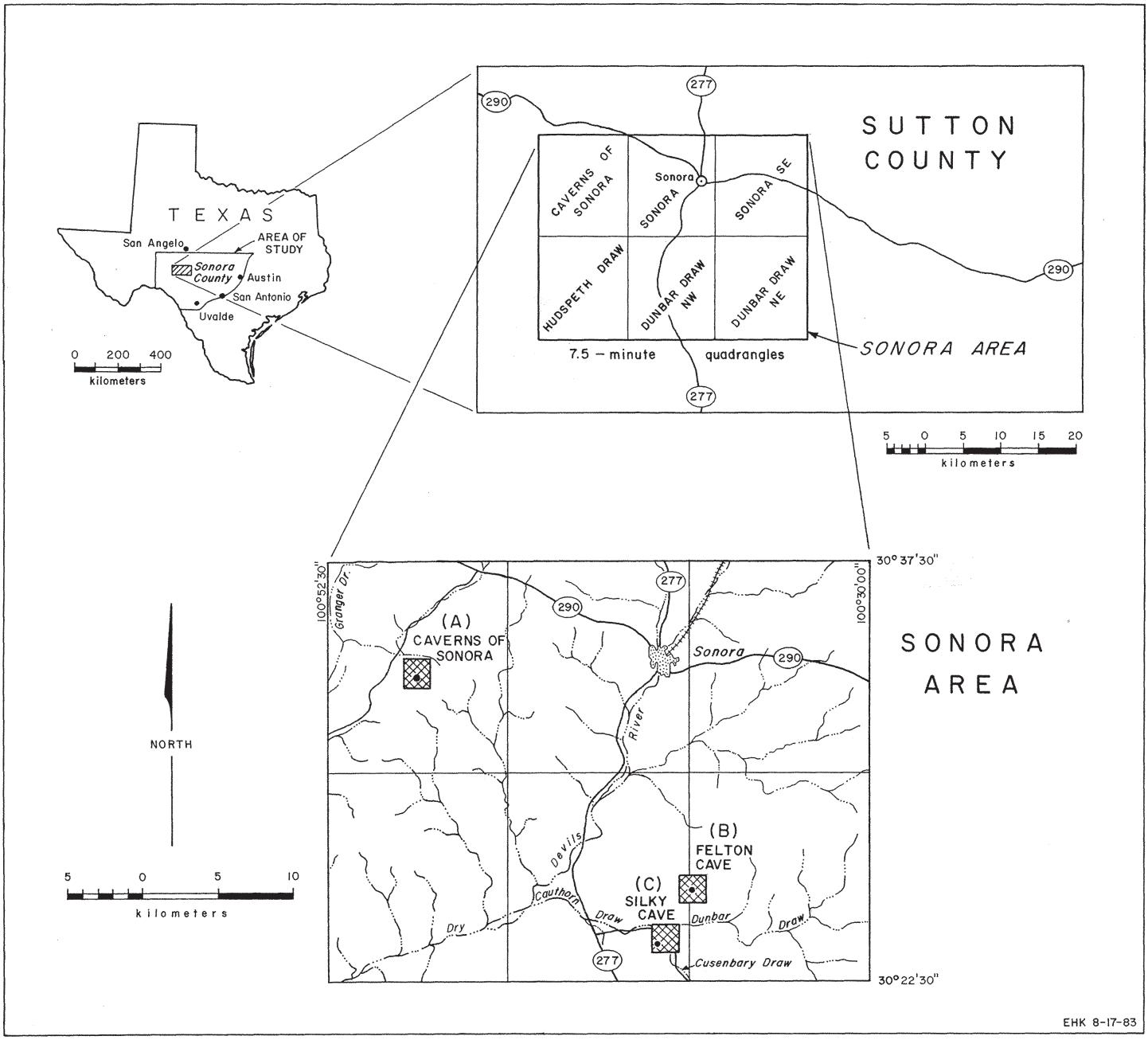


Figure 31: Location map for the Sonora area. Maps showing topography and geology in the vicinities of (A) Caverns of Sonora, (B) Felton Cave, and (C) Silky Cave are shown in Figures 33, 35, and 39 respectively.



Figure 32: Valley of Ice in Caverns of Sonora. Stalactites, stalagmites, and flowstone, as well as walls, ceiling, and floor of the passage are profusely encrusted with cave coral. Note coralloids in lower left corner. Many coral bulbs have grown on faceted crystals (not visible) that developed sub-aqueously at a time when water was ponded in this chamber (see Figure 44). View is to the north. Lighted column is approximately 1.5 m in height.

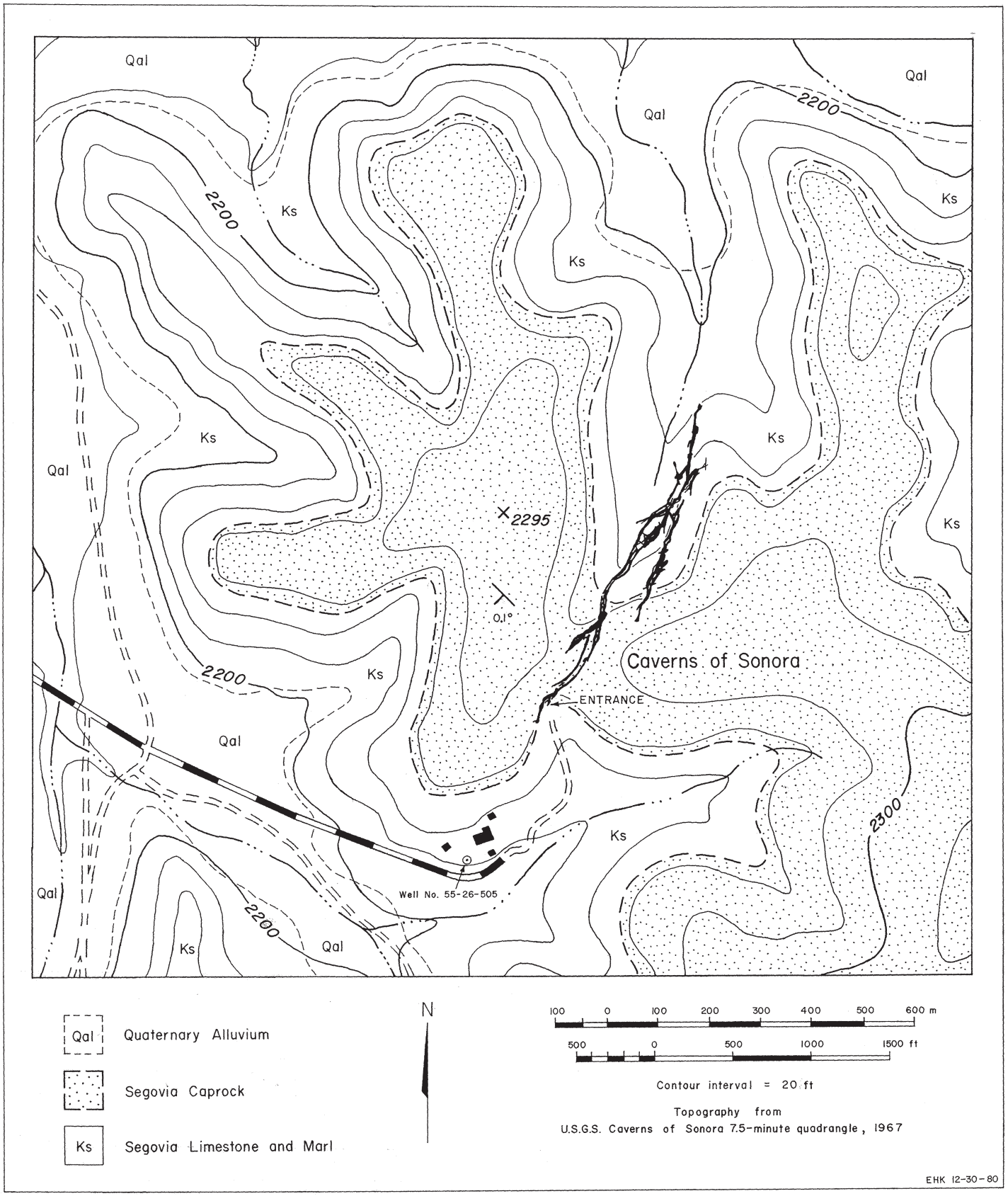


Figure 33: Caverns of Sonora in relation to topography and caprock of the Segovia Formation. See Figure 31 for location and Figure 34 for detailed map of cave. Water-well is from Muller and Pool (1972).

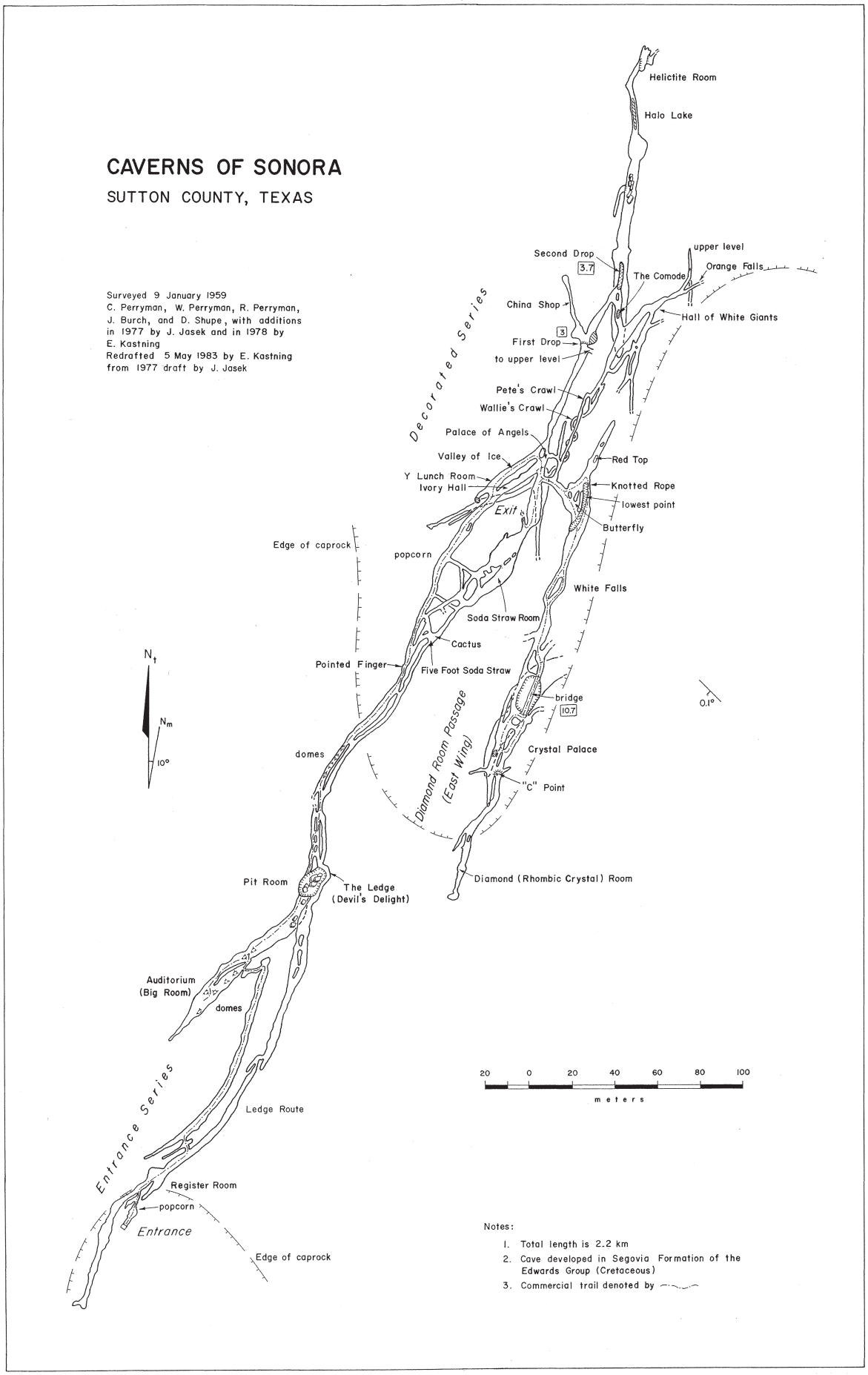


Figure 34: Map of Caverns of Sonora. Compiled from Tandy (1962, figure 1, p. 32), Reddell (1964b, p. 25), and Jasek (1977a, p. 93-94).

Papadakis (1977, p. 303).

A pair of parallel but interconnected passages leads north from the top of the Pit Room. About 100 m from the Pit Room, in the vicinity of the Pointed Finger, passages begin to descend at an inclination of 10° and the character of the cave changes considerably. Up to this point passages have been relatively barren of speleothems, other than cave-coral in the passage leading from the entrance. However, the inner parts of the cave are decorated in such profusion that chambers in the remainder of the cave are unlike any other in North America. Nearly all walls, ceilings, and floors are festooned with delicate helictites, cave-coral, coralloids, soda-straw stalactites, and crystals (Figure 32; see also photographs in Tandy, 1962; Roberts, 1975; Papadakis, 1977; and Kunath, 1980).

The morphology of the cave also changes at this location. The overall configuration of passages is more complex and maze-like in the north part of the cave where there are numerous side passages and multiple levels. Nearly all passages in the cave consist of interconnected, linear segments of varying orientations.

In the rear of the cave upper levels include the Soda Straw Room, Valley of Ice, Palace of Angels, and Hall of White Giants. Lower levels include the Diamond Room Passage, Snowball Alley, and the passage extending from the Second Drop to the Helictite Room (Figure 34). On all levels crawlways form interconnections among the larger passages. The total length of mapped passages north of the Pit Room is presently 1.6 km. It is estimated that this length will be at least doubled once all known passages, many of which are on other levels, are surveyed. Problems of access since the early 1960's have prevented a complete and detailed survey of the cave. My annotations and additions to the map (Figure 34) are from observations and measurements taken while on commercial tours. Additional historical and descriptive information on the cave can be found in Crisman (1956), Gray (1961), Tandy (1962), Reddell (1964b), Price (1973), Halliday (1976), Jasek (1977a), Papadakis (1977), Fieseler and others (1978), Messling and Messling (1978), and Shankle (1979).

The natural entrance of the Caverns of Sonora lies just beneath the lowest resistant bed of caprock of the Segovia Formation (Figures 33 and 34). The upper levels of the cave, in particular the Ledge Route and its continuation north of the Pit Room, have developed within porous, honeycombed beds of the Segovia. Lower levels, including the Auditorium, Diamond Room Passage, and the passage to the Helictite Room, lie within massive, dolomitic, marly units. These lower-level avenues are generally larger in cross section than are those on the upper levels.

In places the beds that form the caprock of the upper Segovia Formation overlying the cave have been stripped by valley erosion, except for a narrow section of ridge overlying passages extending between the natural entrance and the vicinity of the Pointed Finger (Figures 33 and 34). The influences of the caprock on the morphology of passages and on the deposition of speleothems are explained later in this chapter.

The Caverns of Sonora contain no active streams; however, seepage and infiltration occur in the northern sections of the cave. The local, piezometric surface is approximately 40 m below the lowest point in the cave, at an elevation of 597 m above mean sea level (based on (1) wells numbered 55-26-504 and 55-26-505 in table 1, p. 9 of Muller and Pool (1972), (2) figure 5 of Blank and others (1966), and (3) personal communication with Caverns of Sonora personnel in 1978).

It is difficult to determine thicknesses of detrital sediments in the cave because most such deposits have subsequently been covered with flowstone, especially in well-decorated sections. However, based on sediments in the southern part of the cave, deposits composed mainly of sand and silt appear to be less than 1 meter thick. There has been some collapse of ceilings within the cave, and sizeable deposits of breakdown occur in the Pit Room, in the large pit in the Diamond Room Passage, and in the passage leading to the Helictite Room.

The cave passes beneath a topographic divide separating two draws that flow west to Granger Draw (Figures 31 and 33). The lowest passages in the cave are nearly 30 m lower in elevation than the bottoms of these draws.

Felton Cave

Felton Cave, approximately 15.5 km south of Sonora and 3.0 km north-northeast of the confluence of Dunbar and Cusenbary Draws (Figures 31 and 35), is indicated on the Dunbar Draw NE 7.5-minute U.S.G.S. quadrangle. The cave has a surveyed length of 2.05 km, an estimated depth of 25 to 35 m below the entrance, and an end-to-end extent of 350 m (Figure 36).

The single entrance to the cave is a 6 m-wide, 9 m-long, and 12 m-deep pit in a ledge of bedrock on the southern side of a low ridge. A series of climable ledges provides easy access to the cave which trends generally north from the entrance (Figure 36).

At the base of the slope within the entrance an opening on the southern wall leads to the Left Hand Room, a large, nearly circular chamber measuring 35 m wide by 45 m long and 9 to 10 m high. A short side-passage extends south from this room, and several pits are found along its periphery, the deepest of which is 12 m. A short, low crawlway leads northwest form the entrance to a complex series of passages composed of linear segments distributed along several orientations. The maze-like character of numerous intersecting passages characteristic of much of the remainder of Felton Cave. Passages typically range from 2 to 6 m in width and 1 to 3 m in height.

Four major series of passages have been mapped in the cave. The First Series, a maze of short passages averaging 30 m in length extends northwest from the area of the entrance and includes the Subway and Dead Goat Pit. This section communicates at three places with the Second Series, a 300 m-long set of parallel passages striking N 44° E from the Bat Rooms, past Danger Pit, to the Guano Slide. Here crawlways, forming a small maze and averaging 3 m wide and 1 m high, branch to the west where they intersect Silo Pit. From the bottom of this pit the Third Series extends north and south along a strike of N 3° W. To the north it continues as the Breakdown Passage, a series of multi-level, low passages to the 9 m-deep Pink and Purple Pit. Low crawlways continue north from the bottom of this pit. The south segment of the Third Series passes beneath passages of the Second Series (in the vicinity of Danger Pit) to a termination in breakdown. However, 50 m north of o this point and to the east a crawlway leads to the Carlsbad Room, a chamber 70 m in length, up to 15 m in width, and 6 m high. The Fourth Series includes Babcock's Lost Passage which extends N 6° W for 160 m from the southwest end of the Second Series. Near the end of the 1 to 2 m-high passage is a narrow fissure-like passage extending to the west for an undetermined distance.

Nearly all of Felton Cave lies beneath the oldest, resistant bed of caprock of the Segovia Formation (Figures 35 and 36). Except for the southern margin of the Left Hand Room, no passages within the cave are known to extend beyond the margin of the caprock. The upper levels of the cave, including the Left Hand Room and passages of the aforementioned First, Second, and Fourth Series, have developed within porous, honeycombed beds of the Segovia. Lower levels, consisting of passages of the Third Series, lied within underlying, massive, dolomitic, marly beds (Figures 37 and 38).

There are no active streams in Felton Cave, but seepage and infiltration occur near the margins of the caprock, especially in the Left Hand Room, Carlsbad Room, and Pink and Purple Pit. Locally the piezometric surface is approximately 50 m below the lowest levels in the cave (based on well number 55-35-601 in table 1, p. 17 of Muller and Pool, 1972).

Much of Felton Cave is floored with silt and sand. Sediments in the upper level of the Breakdown Passage, in the upper-level passages near Danger Pit, and near the end of Babcock's Lost Passage, locally consist of several centimeters of crystals of calcite (Russell, 1962). Collapse of ceilings has occurred in the Left Hand Room, in the Carlsbad Room, and along the Breakdown Passage, where breakdown blocks up to several meters in size litter floors of passages.

Speleothems are sparse or nonexistent in Felton Cave except in areas of infiltration. The Left Hand Room contains large speleothems including two stalagmites more than 4 m high. Presently, the most active precipitation of calcite occurs as flowstone in the vicinity of Pink and Purple Pit.

Like Caverns of Sonora, Felton Cave passes beneath a topographic ridge (Figures 31 and 35). Its lowest passages lie approximately 30 m lower in elevation than the bottoms of local valleys.

Silky Cave

Silky Cave is 1.71 km southwest of the confluence of Dunbar and Cusenbary Draws (Figures 31 and 39). Its surveyed length is 405 m, total relief within the cave is about 8 m, and the maximum horizontal extent of passages is 200 m from the entrance to the innermost, eastern point (Figure 40).

The only entrance, near the top of a ridge, is a small (1 m-in-diameter) vertical hole in bedrock that drops 1 m into a 15 m-long crawlway. This intersects a small room, 3 m wide, 10 m long, and 2 m high and then continues for an additional 4 m. The cave was enterable only to this point until 1975 when explorers excavated a crawlway connecting to a parallel passage 4 m to the north (Figure 40).

The cave trends generally N 86° E from the entrance to the innermost point and consists of a series of segments of linear passages, giving the cave an angular pattern. There are about a dozen side passages along the main course through the cave, all less than 25 m in length. Most passages vary between 1 and 6 m in width and 0.3 and 3 m in height. Approximately 40 domes or ceiling-pockets are distributed throughout the cave and penetrate as much as 2 m into the ceilings

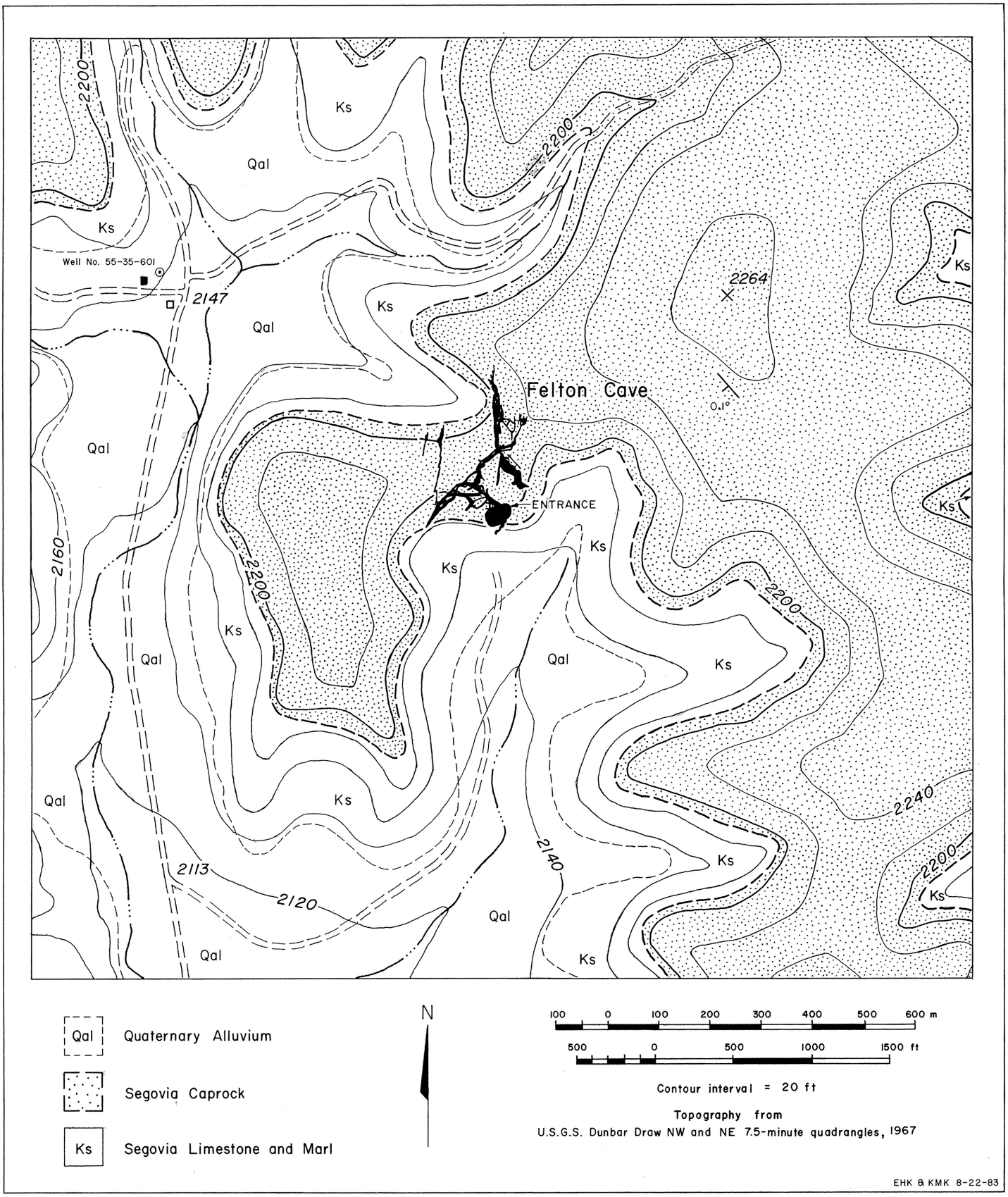


Figure 35: Felton Cave in relation to topography and caprock of the Segovia Formation. See Figure 31 for location and Figure 36 for detailed map of cave. Water-well is from Muller and Pool (1972).

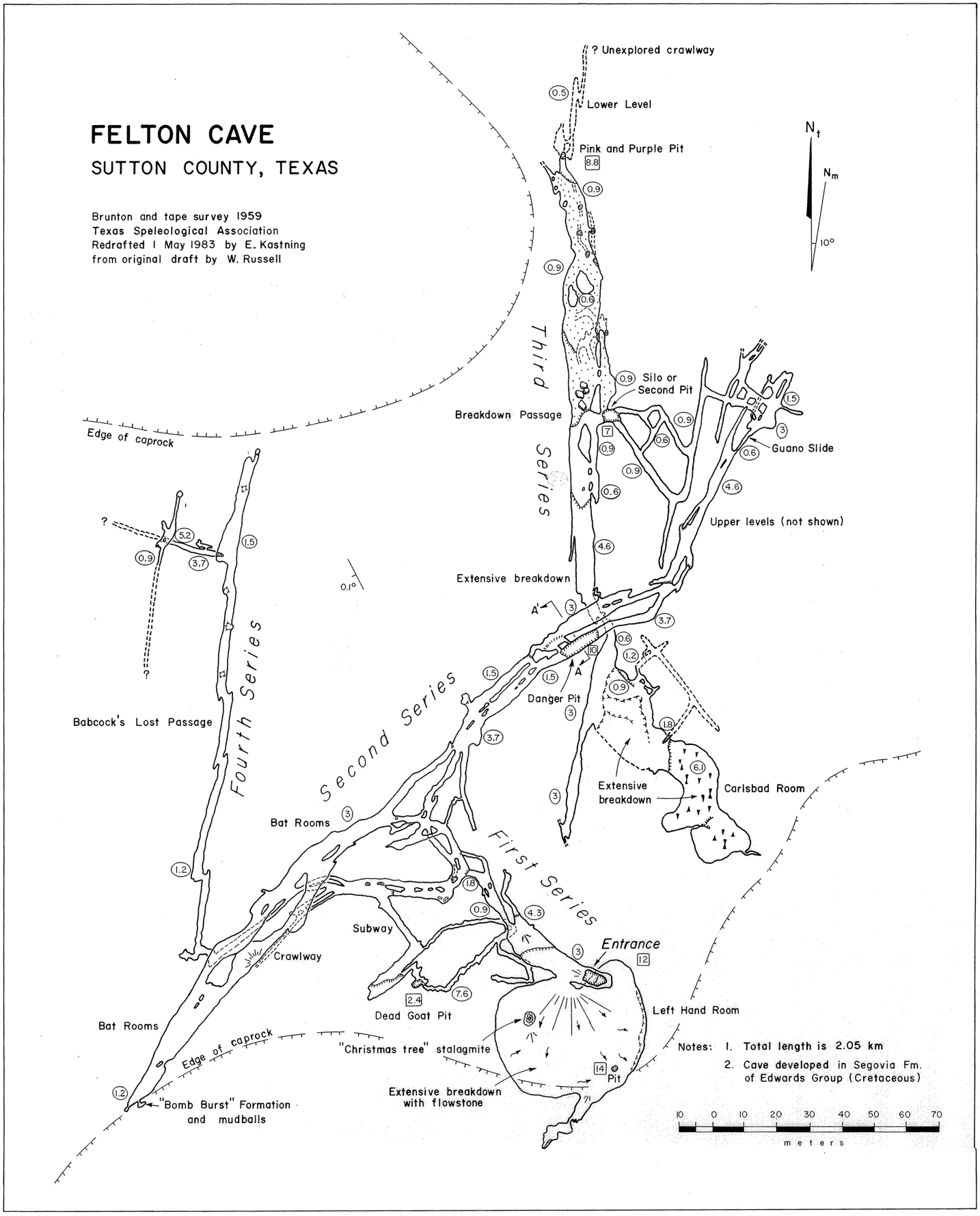


Figure 36: Map of Felton Cave. Modified from Russell (1962). Section A-A' is shown in Figures 37 and 38.

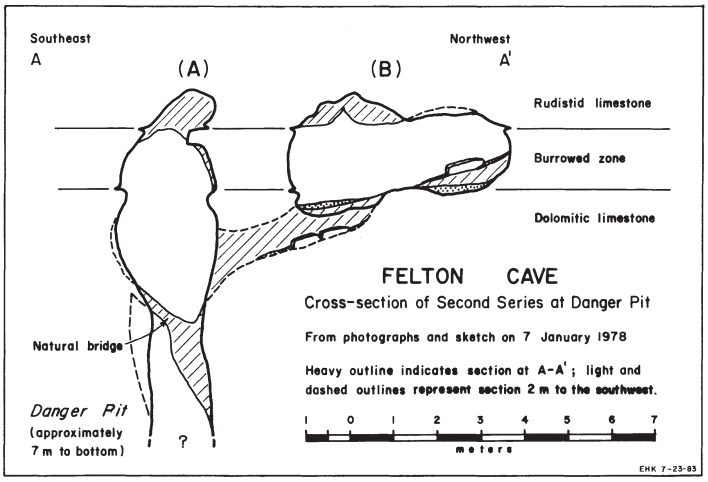


Figure 37: Cross-section of (A) Danger Pit and (B) adjacent passage, Felton Cave (coresponding to section A-A' in Figure 36. See also Figure 38.

(Figure 40).

The entrance of Silky Cave is at the edge of the lowest resistant bed of caprock of the Segovia Formation (Figures 39 and 40). The cave lies totally beneath the caprock and has developed within the porous, honeycombed beds of the Segovia lying just beneath the caprock. Silky Cave is generally very dry and in places very sandy and dusty. Active infiltration is evident only in the innermost chamber where seepage occasionally fills an intermittent pool and has deposited speleothems and dams of travertine. In places along the cave previous infiltration has deposited flowstone, stalactites, and stalagmites, including one 3 m-high column in the western part of the cave. The local piezometric surface is approximately 85 m below the level of the cave (based on wells numbered 55-35-901 and 55-35-901 in table 1, p. 18 of Muller and Pool, 1972).

Much of the cave is floored with silt and sand. Small slabs and flakes of breakdown are strewn on floors along the entire length of the cave; however, larger blocks occur along the central third of the cave, where passages are widest (Figure 40).

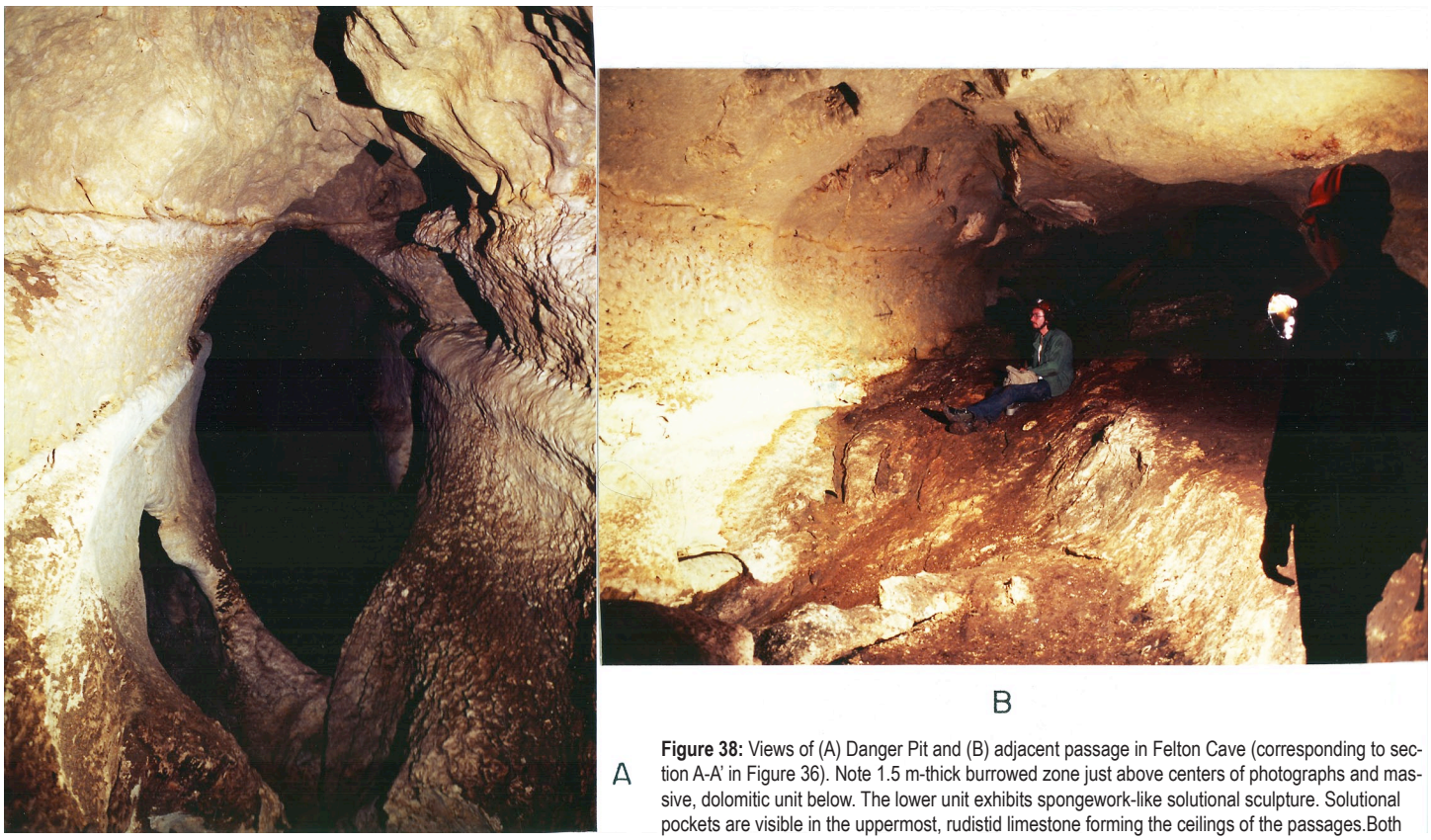
Silky Cave penetrates a relatively flat-toped ridge (Figures 31 and 39). The stratigraphic horizon containing the cave lies approximately 35 to 40 m above the valley of Cusenbary Draw 1.0 km to the north and 1.4 km to the east.

Speleogenesis

The three caves investigated in this area are similar in several ways: (1) they are characterized by horizontal passages along distinct stratigraphic horizons; (2) their passages are angular and consist of interconnected, linear segments; (3) they have developed in the Segovia Formation in porous, honeycombed beds immediately underlying the lowest resistant bed of caprock of that formation; (4) they pierce narrow ridges in a moderately to highly dissected terrain; (5) their lowest passages are presently 40 or more meters above the local, piezometric surface; (6) speleothems are concentrated in passages below the edge of the caprock or in areas over which the caprock has been removed. These similarities suggest that the caves have a common genetic history that has been relatively uniform for most caves in the Sonora area.

Lithologic Control

Most passages of caves of the Sonora area are conformable with the strata that contain them. Upper levels in Caverns of Sonora and Felton Cave, as well as the single level of Silky Cave, have been excavated in part along highly burrowed beds (Figures 37, 38, and 41). In some areas, these passages bifurcate and rejoin, forming a slightly braided or anastomosing pattern (passages A, B, and C or D, E, and F, in Figure 41). This is exemplified by (1) passages extending from the Pit Room to the area of the Palace of Angels and the East Wind of the Caverns of Sonora, (2) the second and third series of passages in Felton Cave, and (3) to a much larger degree some small loops in Silky Cave. Anastomosing tendency is characteristic of development of caves where bedding-plane partings or highly soluble or porous strata provide more efficient flow than do vertically oriented fractures. Moreover, this pattern of development of passages is often a result of periodic flooding (Palmer, 1975). Anastomosing passages in Caverns of Sonora and Felton Cave consist of long and continuous series of short, linear segments (Figures 34 and 36), and scrutiny of ceilings and floors of these passages indicates that most segments are closely guided by joints. These observations suggest that (1) passages were originally excavated in the phreatic zone as single, long conduits with cross sections smaller than those existing today, (2) during a later, vadose phase of development some passages were invaded by free-surface streams which became entrenched into the floors, forming canyons and routes of diversion



A B

Figure 38: Views of (A) Danger Pit and (B) adjacent passage in Felton Cave (corresponding to section A-A' in Figure 36). Note 1.5 m-thick burrowed zone just above centers of photographs and massive, dolomitic unit below. The lower unit exhibits spongework-like solutional sculpture. Solutional pockets are visible in the uppermost, rudistid limestone forming the ceilings of the passages.Both passages exhibit strong control by fractures. See also Figure 37.

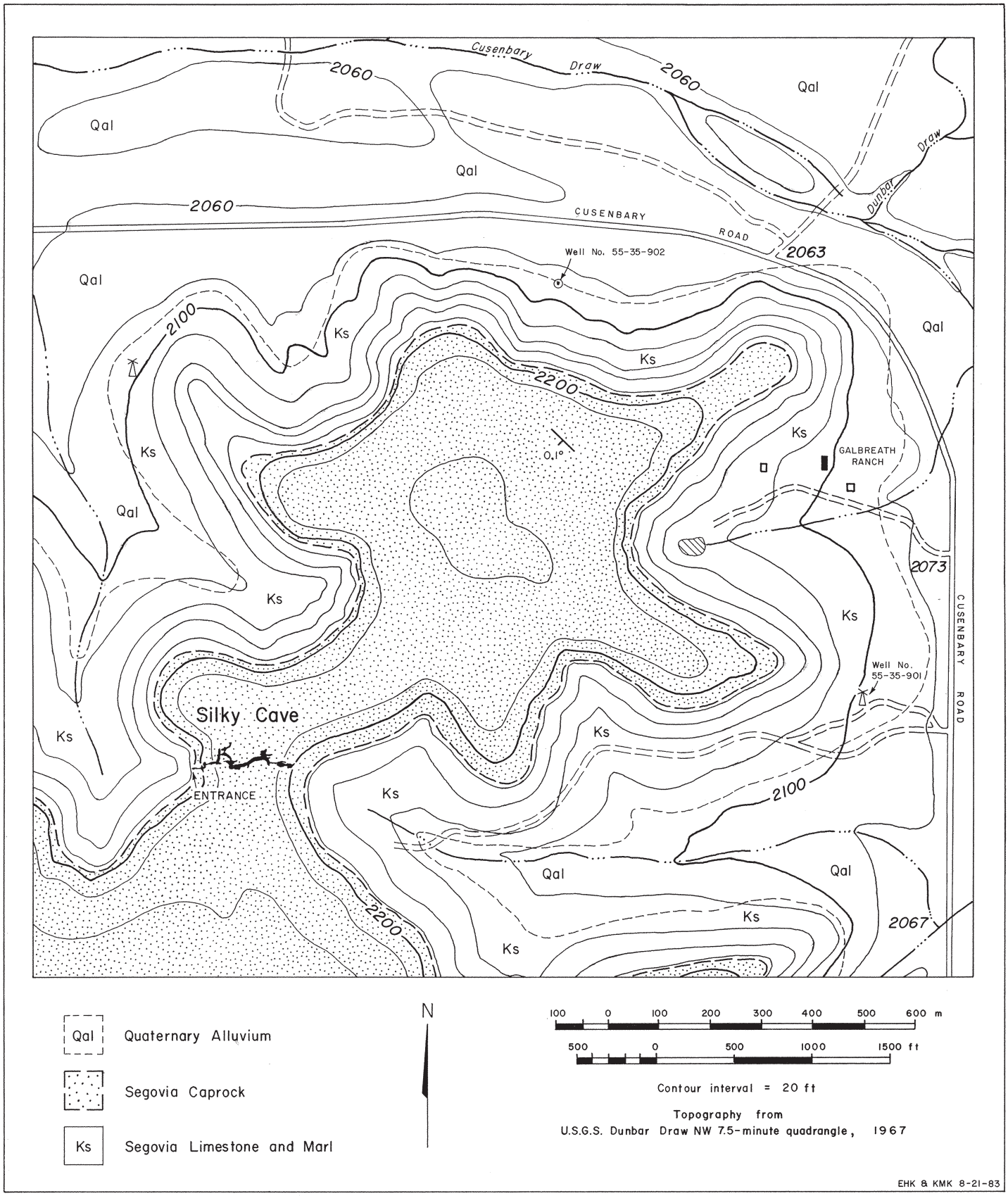


Figure 39: Silky Cave in relation to topography and caprock of the Segovia Formation. See Figure 31 for location and Figure 40 for detailed map of cave. Water-wells are from Muller and Pool (1972).

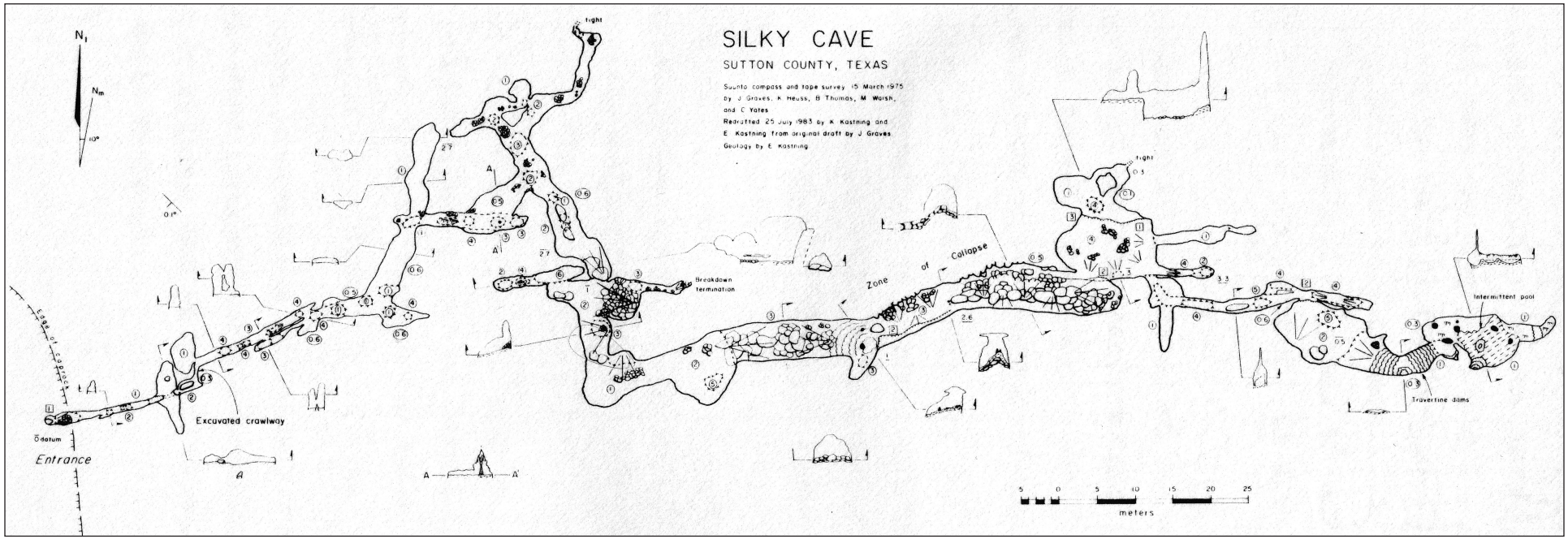


Figure 40: Map of Silky Cave. Modified from unpublished map courtesy of J. Graves.

to lower levels, and (3) those conduits were subsequently invaded and enlarged by floodwaters.

The Segovia Formation in the western part of the Edwards Plateau is less dolomitic and contains less collapse-breccia than does the Segovia of the eastern Edwards Plateau. Indeed, "in Sutton County, the Segovia interval consists mostly of thick beds of rudist and miliclid limestone separated by substantial intervals of ammonite-bearing marl" (Rose, 1972, p. 35). Passages have been enlarged to their greatest volume in beds of marl and limestone that are slightly dolomitic, massive, and consist of finely-recrystallized micrite and fine, skeletal biomicrite or biosparite. Intervening units that are typically more dolomitic, nodular, and cherty are generally less cavernous. However, they do contain short, inclined passages and pits connecting the major passages of other levels (Figure 41).

The lithology of the resistant beds of caprock (Figure 41) has greatly influenced the hydrogeologic mechanisms of speleogenesis and subsequent deposition of speleothems, and variations in lithology have contributed to mechanical failure and collapse of ceiling in lower-level passages. These effects are presented more fully below.

Structural Control

Relationship to Fractures. In each of the three caves over 90 percent of the length of passages consists of linear segments of passages (Figures 34, 36, and 40). Observation has verified that most segments are aligned along fractures visible in ceilings, walls, or floors of passages (Figure 38). Rose diagrams (Figure 42) indicate that orientations of passages for individual caves are clustered, yet they vary widely from cave to cave. Passages of the Caverns of Sonora are clustered in the northeast quadrant, those of Felton Cave are distributed in the northeast and northwest quadrants, and those of Silky Cave lie nearly along an east-west bearing. Cumulatively, passages of all three caves have a dominant north-northeast orientation (Figure 42D). Generally, segments of passages aligned along fractures do not exceed 150 m in length.

There is evidence to suggest that the vertical extent and openness of fractures are limited by the lithology of the bedrock. Orientations of fractures vary with stratigraphic position in two of the caves. In Caverns of Sonora passages developed in the massive, porous beds of the upper levels generally trend from N 20° E to N 55° E, whereas passage in thinly bedded, nodular, and slightly cherty beds of the lower levels trend from N 5° E to N 25° E (Figure 34). In Felton Cave upper-level passages of the Second Series and many of those of the First Series within porous, burrowed beds generally trend from N 25° E to N 50° E, whereas lower-level passages of the Third and Fourth Series in massive,

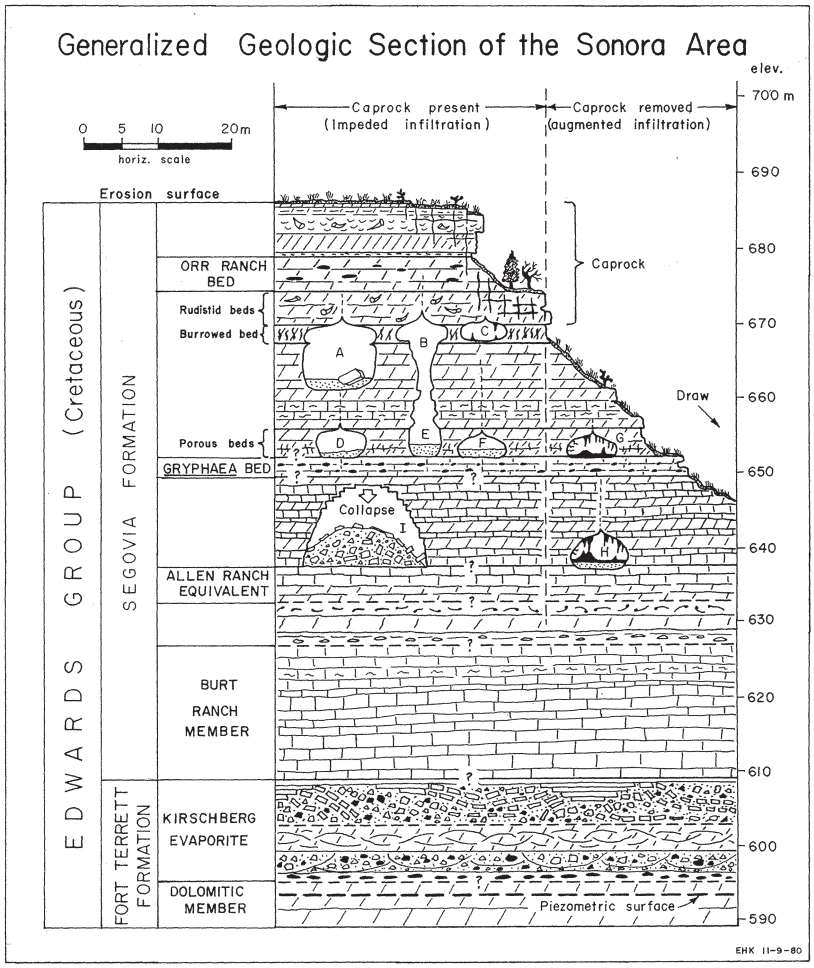
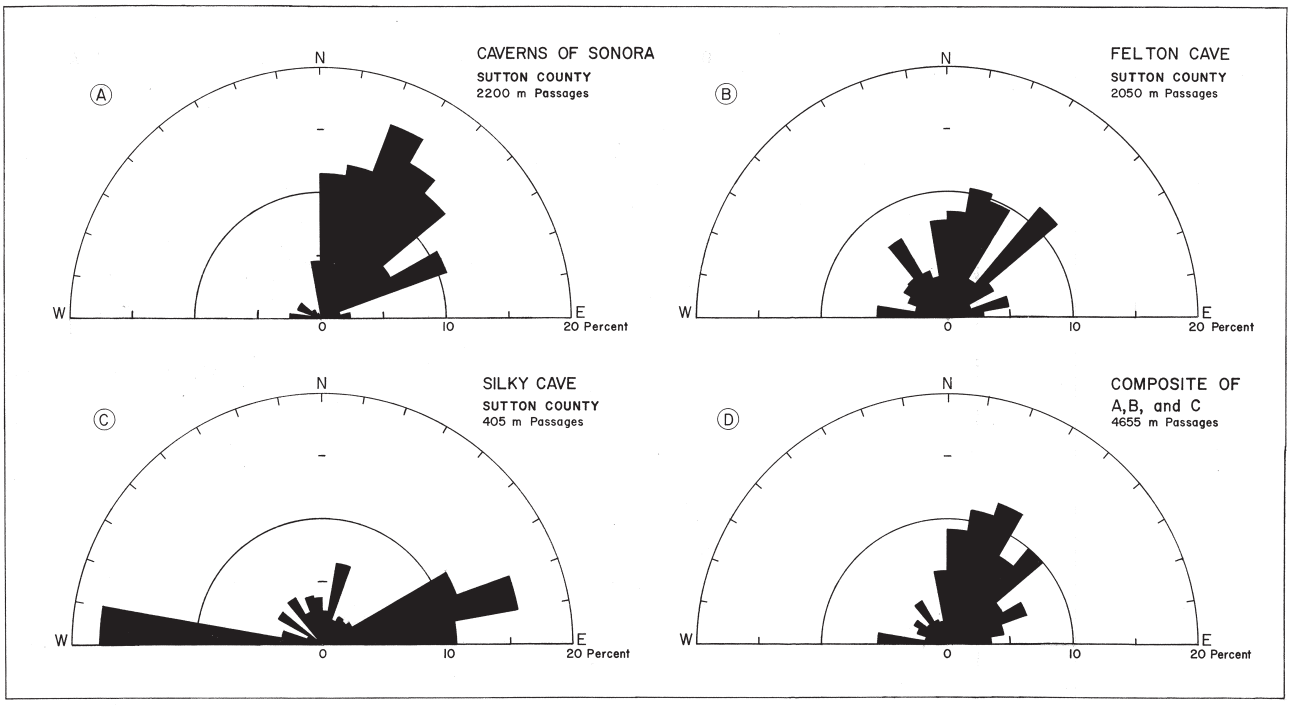


Figure 41: Generalized geologic section of the Sonora area showing stratigraphy and lithology (based on data from Rose (1972, figures 22 and 24) and field-observations), weathering profile, piezometric surface, and horizons of caves. A through I are typical cross-sections of passages. Note proliferation of speleothems beneath areas where the caprock has been removed. Vertical scale varies slightly across area. See also Figure 43.

Figure 42: Rose-diagram of inner segments of passages in (A) Caverns of Sonora, (B) Felton Cave, (C) Silky Cave, and (D) all three caves. Based on measurements from maps of Figures 34, 36, and 40.



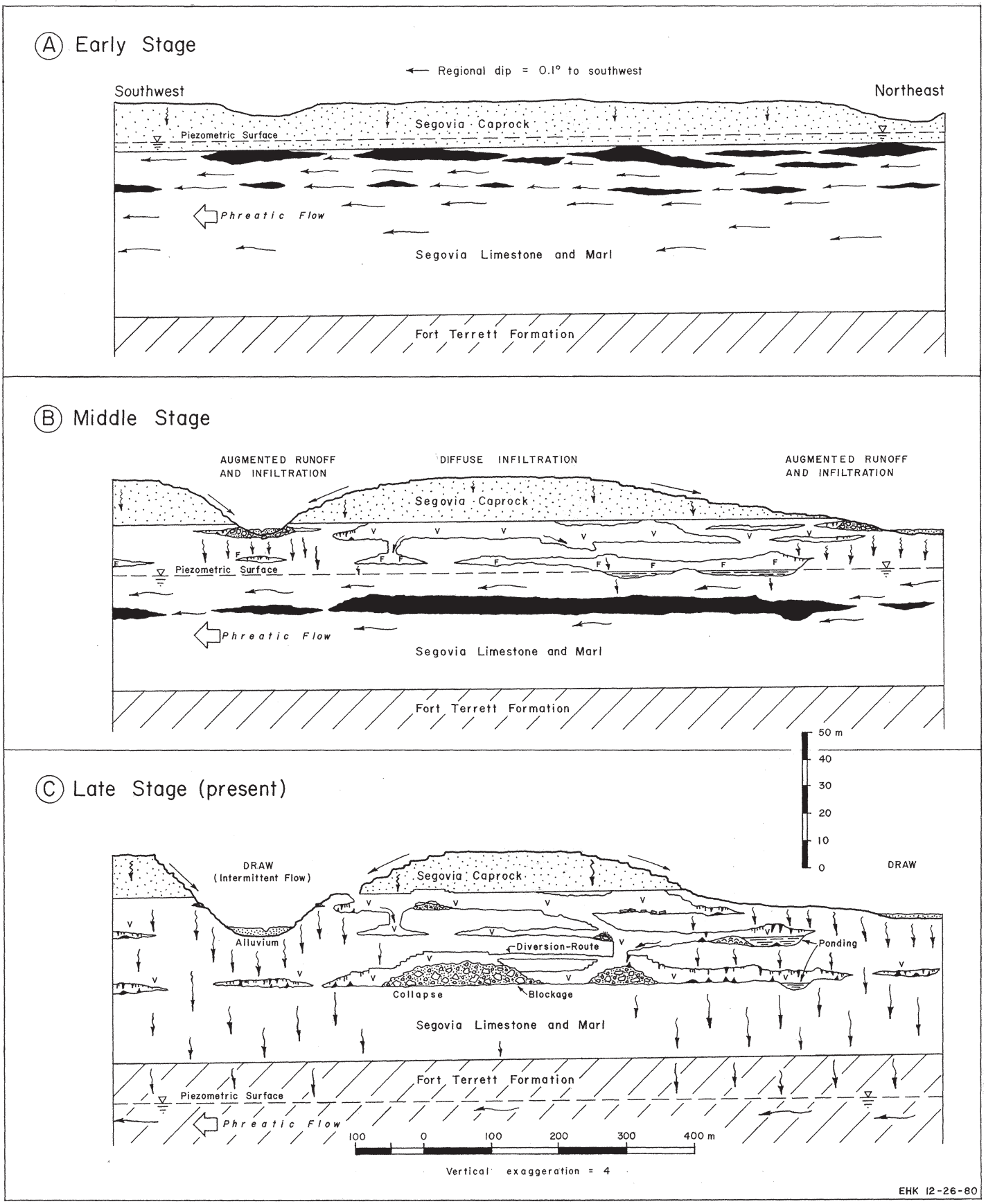


Figure 43: Generalized dip-section in Sonora area depicting sequence of development of caves. (A) Relatively nondissected surface, upper levels of caves undergo enlargement. (B) Baselevels drop, dissection of caprock begins, lower levels of caves develop. (C) Passages are drained, valleys widen, erosion of caprock continues, passages are modified by collapse and stoping, routes of diversion develop around zones of collapse, infiltration and growth of speleothems are augmented. V indicates abandoned passages in the vadose zone. F indicates passages in the floodwater zone. Water-filled passages, shaded in black, are in the phreatic zone.

dolomitic beds trend from N 0° D to N 15° E (Figure 36). In Silky Cave, where all development has occurred in the upper-level zone in massive beds, passages trend from N (Figure 40).

Variations of orientation of passages among levels can be attributed to (1) limitation of the vertical extent of fractures by differences in competency and brittleness of strata, (2) differences in initial openness of fractures among strata prior to dissolutional enlargement, (3) fracturing of lower beds prior to deposition of upper beds, or (4) changes in direction of hydraulic gradients and flow of groundwater as levels evolved in sequence. The first two examples reflect variation in fracturing among strata of differing lithology. From evidence in the three caves upper beds apparently developed fractures of northeastern orientation, whereas lower beds developed north-northeast trending fractures. It is also possible that the lower fractures represent deformation that pre-dates an event responsible for the upper set of fractures. The older fractures may in fact be vertical extensions of fractures form the Fort Chadbourne Fault Zone (Figure 10) of Pennsylvanian age that trends north-northeast through Sutton County (Conselman, 1954, 1958; Rall and Rall, 1958a,b). The upper, younger fractures may be due to extension and unloading, as the overlying Late Cretaceous formations were stripped from much of the Segovia Formation during regional denudation. The higher frequency of open fractures in the upper levels tends to support this possibility. North- and northeast-oriented lineaments may represent tensional fractures caused by differential compaction of Cretaceous strata into Pre-Cretaceous valleys that flank a north-south oriented Pre-Cretaceous topographic high (Cartwright, 1932; Pool, 1976, 1977; see also Figure 10).

Conjugate fractures in the area are generally oriented at right angles to the northern and northeastern sets in Caverns of Sonora and Felton Cave and the eastern set in Silky Cave. However, they have relatively short, horizontal extents, and development of passages along them is much less that that along the primary sets (Figure 42). This may mean that there was little or no tension along the smaller fractures during tectonism, and they remained closed, allowing little, subsequent flow of groundwater. Because they trend subparallel to the strike of the strata rather than along the dip, these fractures may not have been parallel to the direction of flow during enlargement.

It is interesting to compare orientations of fractures in the caves with those of mapped lineaments from aerial photographs (Pool, 1972a, 1977, figures 5 and 6, p. 10-12) and with orientations of elongate playas and other small basins of lakes (Pool, 1973, 1977, figures 11 and 12, p. 14-15). The orientations are predominantly north-northeast, corresponding very well with trends of passages in caves. However, most playas and small lakes, which also have a north-northeast alignment, are not located on lineaments and may not be directly related to increased rates of infiltration and solution along fractures. Instead, the lakes appear to be the product of deflation and solution, but were later modified by erosion from currents and waves over the last 5000 years (Pool, 1973, 1977).

A few surficial lineaments in the vicinity of the caves were visible on 1:20,000-scale photographs in Wiedenfeld and McAndrew (1968) and are indicated on Figure 31. Most surficial lineaments have northeast orientations that coincide well with the predominant orientations of passages of Figure 42.

Modification by Collapse. Collapse of ceilings has occurred in all three caves under study. Collapse has enlarged passages in the central part of Silky Cave (Figure 40), where parallel joints are more numerous and spaced 1 or 2 m apart (perhaps indicating a fault nearby, though none was observed). Here ceilings of passages have been raised approximately 2 to 3 m through stoping, and cross sections are typically arched and angular where blocks have fallen. (The passage is stratigraphically positioned at A in Figure 41, but cross section resembles that of I.) Upward stoping has ceased at the caprock which is more competent than the beds beneath.

Collapse is more pronounced in the Caverns of Sonora and Felton Cave, where stoping has progressed through two or more levels of passages. Upper and lower levels have merged through collapse at the Pit Room and Ten-Meter Pit in Caverns of Sonora (Figure 34) and the Left Hand and Carlsbad Rooms in Felton Cave (Figure 36). In the two areas in Caverns of Sonora the lower passages enlarged to a critical width and height whereupon beds separating them from upper levels gave away, forming chambers up to 15 m high (Passage I in Figure 41 represents a lower-level passage enlarged by upward collapse but not yet merged with passage D above). In this example collapse may have been promoted by intersection of two or more fractures in an area of rapid solutional excavation.

A similar occurrence is seen in the Left Hand Room of Felton Cave. Collapse here has resulted in an elliptically shaped (in plan view) room with a wide cone of debris forming its floor. It is interesting to note that if the fractures of the Third Series of passages on the lower level are extended southward, they intersect the Left Hand Room at its apex and pass through the point where collapse has progressed to the surface and formed the doline at the entrance. This suggests that the Left Hand Room has formed at the intersection of major fractures, in particular those of the First and Third Series passages. This example is comparable to large-scale collapse-structures at Natural Bridge Caverns, Comal County (Chapter 10) and Devils Sinkhole, Edwards County. In contrast to the uniform collapse and elliptical shape of the Left Hand Room, beds of the ceiling in the Carlsbad Room have become dislodged along north-south and northwest-southeast sets of fractures, giving rise to an irregular and angular room (in plan view). Probably one large zone of collapse extends on the lower level from the Carlsbad Room to the Left Hand Room, and although no passages have yet been found between these areas, in all likelihood the drainage from the Third series once extended south at least as far as the Left Hand Room.

Zones of collapse and stoping in Caverns of Sonora and Felton Cave lie close to the edge of the caprock, where increases in infiltration may have enhanced solutional enlargement of passages and weakening of bedrock, which in turn prompted collapse of overlying rocks. Collapse may have been a widespread and continuing process, ultimately reaching the surface, and it may have contributed significantly to local removal of caprock and retreat of slopes of valleys.

Hydrogeologic and Topographic Controls

Concordance of levels in caves with bedding and orientation of longer segments of passages in a northeastern direction suggests that groundwater flowed down the very gentle (0.1) southwest dip during speleogenesis. This is supported by the equally gentle southwest dip of the present, regional, piezometric surface (Peckham, 1963; Blank and others, 1966; Muller and Pool, 1972; Walker, 1979). Presumably, the direction of regional movement of groundwater remained unaltered during denudation and dissection of this part of the Edwards Plateau, and the only significant change in the piezometric surface has been a lowering in response to changes in baselevel.

Recharge to the Segovia Formation from the surface of the plateau has been through diffuse infiltration rather than at discrete inputs. This is suggested by (1) lack of dolines or swallow holes on the surface, (20 low density of open fractures in the resistant caprock, and (3) patterns of passages forming networks, spongework, and anastomoses—in contrast to dendritic patterns typical of discrete recharge (Palmer, 1975). The caprock, although hard and relatively resistant to erosion, is nonetheless brittle. It is fractures, but joints are relatively closed and widely spaced. Vertical percolation through the caprock has been slow and uniformly transmitted to the massive, porous beds beneath. In the initial stage of speleogenesis, fractures beneath the caprock may have enlarged uniformly in various sectors of the caves, giving the initial impetus towards development of mazes of the network or anastamoses varieties.

During the early stages of groundwater flow, the surface of the plateau consisted of wide areas of relatively nondisssected terrain (Figure 43A). However, the surface of the plateaus was subsequently dissected by lowering of regional baselevel and incision of major streams (Figure 43B). Channels on the surface breached the caprock and the topography evolved into a complex of narrow ridges which now form divides within a highly dendritic drainage net. This has drastically altered recharge to the limestones beneath. Instead of slow infiltrating into the surface of the plateau, precipitation falling on the uplands contributes to runoff, flowing rapidly down steep sides of valleys to nearby streams.

As the piezometric surface dropped within the caprock, the massive and permeable rocks lying just beneath the caprock became the dominant horizon for the flow of groundwater which moved down the hydraulic gradient to the southwest (Figure 43B). Presumably, recharge to this system of flow was provided by diffuse infiltration into the less dissected plateau to the northeast.

Lower levels, such as the route to the Helictite Room, Diamond Room Passage, and Auditorium in Caverns of Sonora and passages of the Third and Fourth Series in Felton Cave, may have been enlarged as the piezometric surface continued to drop. Eventually, as levels of water fell further, passages were drained, and today the piezometric surface lies within the Fort Terrett Formation, at least 40 meters beneath the lowest known passages in these caves.

Discharge of groundwater in the Sonora region occurred along springs in the evolving drainage basins to the south and southwest. These outlets may at times have included Juno (Beaver Lake), Pecan, Hudspeth, Huffstutter, Finegan, Dolan, Gillis and Slaughter Bend Springs along the Devils River in Val Verde County, which lie at elevations between 335 and 525 m above mean-sea-level (Blank and others, 1966; Reeves and Small, 1973; Brune, 1975). Underflow may

have continued further to the southwest and emerged at springs along the Rio Grande, including Goodenough, McKee, San Filipe, and Cantu Springs (Reeves and Small, 1973; Brune, 1975). Moreover, some discharge may also have been to the south and east, through numerous springs within the Nueces River Basin in Edwards, Real, and Kinney Counties, including Kickapoo, Mud, Pinto, and Los Moras Springs (Long, 1958, 1962, 1963; Bennett and Sayre, 1962; Blank and others, 1966; Brune, 1975).

There is evidence that groundwater flowing to the springs is confined to the permeable, dolomitic member of the Fort Terrett Formation (Figures 41 and 43C), where it is under slight artesian pressure. South of the Sonora area, water rises in wells that intersect the dolomitic aquifer and many of the springs are artesian (Blank and others, 1966). This suggests that more effective aquicludes exist there than in the Sonora area. Recharge to the artesian system is principally through walls and floors of valleys in uncapped regions north of the springs, where rainfall is absorbed and then flows along vertical fractures and solutional channels (Figure 43C). Much of the surficial runoff from infrequent, heavy storms disappears into the alluvium of fluvial channels which consist of a mixture of soil, clay, and gravel, cobbles, and boulders of limestone. Some of this infiltrated water undoubtedly becomes groundwater within the alluvial aquifers, but the remainder is carried below in the carbonate aquifers.

Maze-like configurations of passages in the three caves indicate some specific, hydrogeological controls on speleogenesis. Although their initial development may have been imparted by diffuse infiltration through a caprock to a well-jointed, permeable rock beneath (Palmer, 1975), the position of mazes with respect to other cavernous features suggest that their development may be a later modification in morphology. The complex mazes of the First Series and the north end of the Second Series in Felton Cave (Figure 36) may be products of flooding and ponding of water behind zones of collapse in the Left Hand and Carlsbad Rooms. These tube-like conduits of relatively small dimension may have served as diversion channels around constrictions. They exhibit variable cross-sections, joint-spurs, and abrupt terminations in bedrock, all features common to mazes developed by floodwater (Palmer, 1975). Similar but less extensive areas of maze-work in Caverns of Sonora, in the vicinities of the Soda Straw Room, Palace of Angels, Hall of White Giants, and Crystal Palace (Figure 34) may also have formed in response to collapse-induced flooding near the edge of the caprock, especially at the Pit Room and Ten-Meter Pit. Abruptly terminating, fracture-controlled passages in Silky Cave (Figure 40) indicate that development of a network-like maze by floodwater progressed for a brief period, perhaps in response to collapse in passages west of the entrance, now destroyed by retreat of the slope of the valley. (See Palmer [1975]) for a classification of caves according to morphology [e.g. network, branchwork, anastomoses] and a discussion on their origin.)

It is interesting to note that the three caves studies in the Sonora area pass beneath narrow necks of the ridges they occupy. In all three the entrances are at the down-dip edge of the caprock and the rear sections of the caves lie beneath or extend beyond the margin of the caprock on the up-dip edge of the caprock (Figures 31, 33, 34, 35, 36, 39, and 40). Perhaps the caves functioned as subterranean cutoffs during the early stages of topographic dissection. Hydraulic gradients along fractures in beds immediately beneath the caprock may have been oversteepened where one valley along a ridge in question has been incised to a lower elevation than has the valley on the opposite side. Such an increase in hydraulic gradient accompanied by a concentration of recharge would have promoted large discharges of groundwater through the fractures, which in turn accelerated dissolution and enlargement of passages. Ultimately, if baselevels had not dropped as rapidly as they did, the caves would have weakened the narrow areas of the ridges, and surficial erosion would have destroyed the caves, creating artificial cutoff-channels in the process. This argument can be taken one step further by postulating that some, if not many, present-day positions and orientations of draws may have been determined by subterranean piracy of prior surficial streams. Where such diversions may have occurred, most pre-existing caves have since been obliterated by incision of subsequent streams. This analysis is speculative and is based in part on the positions of major caves beneath narrow segments of ridges in the Sonora area.

Sediments and Speleothems

Deposits of sediments on the floors of caves in the Sonora area are thin in comparison with those of other caves on the Edwards Plateau. Much of the residuum derived from dissolution of the carbonates and sediment washing in from the surface has been flushed form the caves. This may have been accomplished by an increase in discharge of groundwater during cold and wet climates, through flow beneath capped ridges in accordance with subterranean piracy, or both.

Speleothemic deposits are concentrated in the vicinity of the edge of the caprock or where passages extend beyond the caprock (Figures 41 and 43C). Here the permeability of the bedrock and infiltration of meteroic water have been enhanced by removal of the caprock and by fracturing of its outer edge during incipient retreat of slopes of valleys. Passages of the northern two-thirds of Caverns of Sonora, which lie beyond the caprock (Figure 33), have been profusely decorated and are now very moist, in contrast to passages in the southern third of the cave (compare photographs in Tandy [1960, 1962] and Roberts [1975]). Continuous seepage in uncapped parts of the cave has produced a profusion of helictites (Figure 32). Helictites require sufficient moisture and release of carbon dioxide from solution for evaporation and crystallization to occur at their tips, but on the other hand, flow of water through the small central canal within each helictite must be slow enough to preclude the formation of drip and flow deposits (Prinz, 1908; Moore, 1954; W. B. White, 1976). Presumably, permeability of the uncapped limestone at the Caverns of Sonora has been such to have allowed sufficient but not excessive infiltration, in quantities compatible with available rainfall in the area.

In contrast to Caverns of Sonora, Felton and Silky Caves lack extensive, helictite deposits because their passages lie wholly beneath the caprock where infiltration has been impeded. However, rapid seepage just within the edge of the caprock has produced drip- or flow-deposited speleothems (e.g. stalactites, stalagmites, flowstone), many of which are no longer actively forming.

Deposits of helictites, which must post-date the removal of the overlying caprock, is an active process today. Although not part of this study, radioisotopic dating of the oldest helictites would yield a minimum age for the breaching of the caprock over the cave of the incision of valleys and retreat of slopes.

Other speleothems in Caverns of Sonora give evidence for reflooding of large areas of the cave. An extensive overgrowth of calcite covers walls, ceilings, floors, and many deposits of flowstone and dripstone (Figure 32). Encrustations generally consist of projecting, faceted crystals of monocrystalline calcite (Tandy, 1962, figure 7, p. 36; Kunath, 1980, photograph, p. 77). Such forms are characteristic of sub-aqueous crystallization from solutions that are supersaturated with respect to calcite and imply that passages must have been flooded to levels above that of encrustations (W.B. White, 1976). Monocrystalline stalactites and stalagmites (Halliday, 1959) exhibiting hexagonal cross sections and large rhombohedral crystals in the Diamond Room (Quinlan, 1959, figure 5, p. 4; Tandy, 1962, figure 6; Kunath, 1975, 1980, photograph, p. 17) may also have formed in part in a sub-aqueous environment.

Botryoidal overgrowths of calcite, known variously as cave coral, cave popcorn, or cave grapes, are also abundant in many passages of Caverns of Sonora (Figure 32; see also photographs in Girard, 1964, p. 19, and Roberts, 1975). Individual bulbs range from less than 1 cm (globulites) to more than 4 cm in diameter. Cave coral represents a progressive outward growth of calcite as thin films of moisture evaporate or as carbon dioxide is released from solution, and growth commonly originates from small projections on walls, ceilings, and floors of caves, or on other formations in a sub-aerial environment (W.B. White, 1976). In Caverns of Sonora many botryoidal deposits occur on projecting, faceted crystals that have previously formed under sub-aqueous conditions. The best examples of this phenomenon are numerous encrusted stalagmites (coralloids) throughout the cave (see photographs in Crisman, 1956, p. 33; Roberts, 1975, cover, p. 14, 19; Jasek, 1975b, 1979; Johnston, 1979; and Kunath, 1980, p. 17). They exhibit three distinct episodes of deposition; in order these are: (1) the original stalagmite produced by vertical drippage, (2) faceted crystals of calcite produced sub-aqueously during reflooding, and (3) cave coral produced by thin films of moisture in a sub-aerial environment (Figure 44). This sequence dictates that at least parts of the cave experienced (1) an initial drainage of phreatic water and deposition of speleothems by dripping and flowing water, (2) reflooding by supersaturated water and deposition of calcitic spar, and (3) cessation of flooding and depositions of cave coral.

Because helictites, cave coral, and deposits formed by dripping and flowing are all actively forming today within some chambers of the cave, it is apparent that speleothems thrive within the range of moisture provided by seepage through the uncapped Segovia beds. This is in marked contrast to the drier, capped, southern third of the cave where few but desiccated speleothems are disintegrating them (Burch, 1967).

Apparently only the lowest levels of Caverns of Sonora flooded during sub-aqueous deposition of calcite. Ponding may have occurred where vertical drainage may have been blocked by sediment and flowstone on floors or where lateral runoff was impeded by dams of flowstone. Under such conditions, inflow of meteroic water exceeded drainage, thus maintaining pools over a considerable time. Ponding may have occurred during cold and wet, pluvial climates that possibly existed in Texas

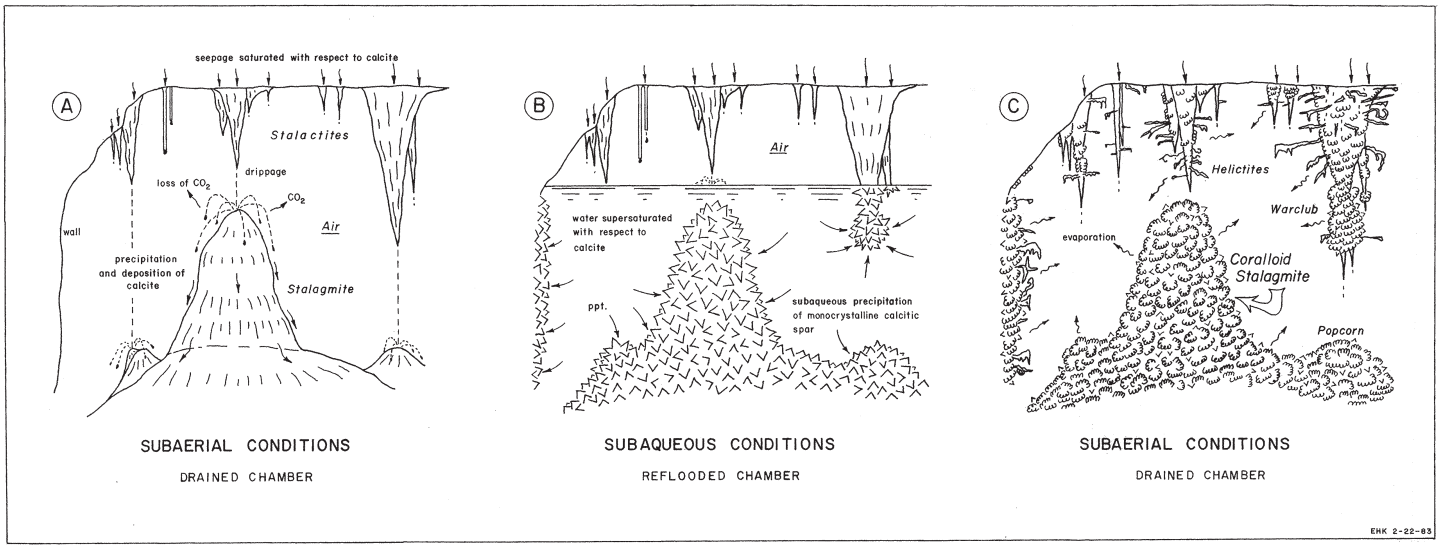


Figure 44: Evolution of a corraloid stalagmite. (A) Stalagmitic accumulation through drippage and splash under sub-aerial conditions. (B) Encrustation of monocrystalline calcitic spar under sub-aqueous (reflooded) conditions. (C) Accumulation of cave coral from thin films of moisture under sub-aerial conditions.

during periods of Pleistocene glaciation in the northern United States. However, supersaturated waters capable of producing sub-aqueous crystals of calcite generally are much warmer than those which may have been introduced by wetter, pluvial climates. As seen in later chapters, most speleothems within caves on the Edwards Plateau that are known to have flooded have been redissolved rather than deposited during reflooding. Consequently, ponding in Caverns of Sonora may have been a Holocene event rather than one of the Pleistocene. This problem might possibly be resolved by isotopic dating of monocrystalline calcite of sub-aqueous origin.

Chronology of Speleogenesis

The following chronological sequence is proposed for the origin of caves in the Sonora area:

1. The Segovia Formation was fractured under three separate processes: (1) differential compaction of these sediments over a north-south-trending structural high, (2) upward extension of pre-existing fractures related to the Chadbourne Fault System of Pennsylvanian age, or (3) unloading through erosion of overlying Upper Cretaceous rocks. The first two of these process produced fractures dominantly oriented north or slightly to the northeast. The latter event may have been responsible for fractures of most other orientations. Vertical continuity of fractures varies among distinct lithic units, with the greatest frequency of fractures occurring within the burrowed and massive units of dolostone beneath the dense beds of caprock, and within underlying, thin-bedded, dolomitic units of limestone of the Segovia Formation (Figure 41).

2. Groundwater began to flow down the shallow, regional dip to the southwest. The piezometric surface was initially within the caprock, but the greatest rates o flow occurred in the uppermost beds of the more permeable rocks beneath. Upper passages enlarged along those routs of flow that followed fractures which were oriented along the dip and were originally mostly open. Infiltration directly through the caprock was slow and diffuse and many have initiated development of anastomotic and network-like mazes in the upper levels (Figure 43A).

3. Regional baselevels began to drop as the Devils River cut its channel to the west and south. Local erosion of upper Segovia units followed and zones of increased infiltration evolved where surficial streams cut deepest into the caprock (Figure 43B).

4. As local baselevels continued to descend, hydraulic gradients steepened and flowrates increased. Passages of larger dimensions began to develop on lower levels in response to a lowering of the piezometric surface and increases in discharge through porous and fractured beds. Valleys breached the caprock and some of the better developed, subterranean flowpaths may have captured surficial runoff. Such pirated streams may have been diverted beneath surficial drainage-divides into he groundwater of adjacent basins. Much of the underglow generally followed the hydraulic gradient in concordance with the regional dip to the southwest.

5. Sections of lower-level conduits enlarge to dimensions where ceilings became unstable and collapsed. This was particularly pronounced near the edge of the caprock where enlargement of valleys further weakened the bedrock. In places lower-level passages intersected overlying passages through stoping. Massive collapse was promoted where two or more prominent joints intersected. Some zones of collapse blocked active, phreatic conduits and resulted in ponding and backflooding in existing passages. This led to development of maze-like passages and enlargement of routes of diversion which eventually provided avenues for floodwater past some of the constrictions (Figure 43C).

6. Piezometric levels dropped beneath passages at the lowest levels and into the Fort Terrett Formation (Figure 43C). Deposition of dripstone and flowstone began as the caves were drained, and further collapse occurred in areas where the caprock receded from above the cave. In some localities where the caprock had been removed, and where underlying caverns had weakened the rock, ridges may have been breached by surficial erosion, allowing streams on the surface to be diverted into neighboring basins. Such cutoff and piracy may have been fundamental to the development of outlying hills and the angular, dissected drainage in this part of the Edwards Plateau.

7. Recharge to the caves increased during the Pleistocene, when this region of Texas experienced heavier precipitation and reduced evaporation. Sections of many caves flooded during storms that were more frequent and intense than they are today. As a consequence, many passages, including those in mazes, may have been substantially enlarged and extended.

8. As milder climates ensued during the late Pleistocene and Holocene, recession of the edge of the caprock continued and infiltration increased over segments of the caves that were no longer protected by the caprock. Dripstone and flowstone were deposited in these sections, and in places these may have created barriers behind which water later ponded as seepage increased. Some small and shallow pools formed behind dams of rimstone, but other, larger and deeper pools were contained by massive deposits of flowstone and collapsed debris. Deposits of crystals of calcite were precipitate below the surface of the water within pools, and covered many surfaces of passages and older speleothems. Subsequently, reduced rainfall or breached dams allowed the pools to drain. Continued slow seepage into passages underlying the sides of valleys promoted a profuse, secondary growth of helictites and cave coral.

SUMMARY

The Sonora Area is representative of the dissected, upland surface of the Edwards Plateau, where extensive caves have developed beneath crests of ridges and in erosional outliers. From the Sonora examples, it is apparent that speleogenesis has progressed from upper levels to lower levels in response to lowering of baselevels and the piezometric surface. Regionally, groundwater has continually migrated down the dip to the southwest. Passages have been modified at various stages by collapse and deposition of speleothems, in response to erosional removal of a resistant caprock. The caves are relict features now considerably above floors of nearby valleys, yet they represent moderately well integrated systems of flow that evolved early during the erosion and dissection of the plateau.

7
MENARD AREA

DESCRIPTION OF AREA

The Menard area lies within Menard County, approximately 12 km west of Menard, the county seat. It is the northwesternmost subarea investigated on the Edwards Plateau, and represents the limit of extensive karst development in that direction (Plate 1). The longest cave in Texas, the Powell Cave System (more than 20 km of mapped passages), is located here. Despite extensive outcrops of Cretaceous limestone and moderate, topographic relief, all other known caves in the county are less than 50 m in length (Appendix C). The Powell Cave System may potentially be 6 to 8 times longer than any other cave of the Edwards Plateau. The great length and extent of the system, in a region of otherwise poor development of caves, implies that locally favorable geologic and hydrologic settings have significantly augmented speleogenesis within a relatively small area.

The Menard area, as chosen for this study, is bounded on the south and north along lines of 30°52' and 30°58' north latitude respectively, and on the east and west along lines of 99°50' and 100°00' west longitude respectively (Figure 45). The San Saba River drains the entire area.

Baker and others (1965), Mount and others (1967), and Barnes (1981a) have mapped the geology of the area at scales ranging from 1:250,000 to 1:500,000. For greater detail I have mapped the geology of the bedrock of the Menard area to a scale of 1:20,000 from aerial photographs published in Coffee and others (1967) and have used the stratigraphic nomenclature of Barnes (1981a). Figure 45 shows the distribution of rocks in outcrop and includes major caves and springs of the area.

The Fort Terrett and Segovia Formations of the Edwards Group underlie nearly the entire area. A small outcrop of Permian limestone is visible at Beyer Crossing, where the San Saba River has cut down below the Cretaceous. Quaternary alluvium occurs along the San Saba River and some of the major creeks.

The Fort Terrett Formation is exposed in valleys of rivers and creeks, such as those of the San Saba River and Rock, Clear, and Dry Creeks. The upper part of the Fort Terrett is a porcelaneous limestone that grades downward into a cherty, medium- to course-grained, massively to thinly bedded limestone. Below this it becomes more dolomitic and passes through some molluscan-fragment beds into extensively burrowed units. The lowermost beds are marl or marly limestone.

The Segovia Formation forms the uplands. It consists of a sequence of dolomitic units of limestone and marl. The lower part is a marly limestone comprised of thin to massive beds which are in part nodular and more resistant than the underlying Fort Terrett Formation. Consequently, the lower Segovia Formation weathers to steeper slopes than does the upper Fort Terrett Formation.

The Cretaceous rocks of the area lie generally parallel to the Permian-Cretaceous unconformity (Cartwright, 1932, figures 1 and 3; Barnes, 1944, p. 35, 42-43; Mount and others, 1967, plate 10). Geologic mapping from aerial photographs and measurements in the field indicate that Cretaceous beds strike N 47° E and dip approximately 0.18° to the southeast. Although beds are nearly flat-lying, this very shallow dip has exerted significant control on the development of caves.

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

Powell Cave (also known as Jack Pit Cave, Mushroom Cave, and Railroad Trap), Neel Cave, and Silver Mine Cave are collectively known as the Powell Cave System. The entrance to Powell Cave is 2.75 km north of Beyer Crossing on the San Saba River. The entrances to Neel Cave and Silver Mine Cave are located 1.55 km to the southeast and 1.65 km to the northwest of the Powell Cave entrance respectively (Figure 45). Lakeside Spring, the resurgence for streams within these caves, is located 2.40 km east-southeast of the entrance to Neel Cave. A collapsed doline, known as Meteor Hole, is located approximately 2.30 km northwest of Silver Mine Cave along Dry Creek (Blakemore, 1939), but it has since become filled (Lindsley, 1967b).

Powell Cave had been entered and partially explored in the early part of this century (a 1916 date has been found cut into a wall). However, it was not until the early 1960's that the cave came to the attention of speleologists. Exploration and mapping of the cave began in 1962 and continued through 1968 as a series of five projects of the Texas Speleological Association (anonymous, 1962c, 1964a,b,c; Lindsley, 1963a, 1967a,b,c, 1968; Estes, 1967; Russell and Reddell, 1968). Approximately 16.2 km of passages were surveyed during these efforts. Interest in the Powell Cave System was renewed in 1976. Several additional projects have added at least 1.2 km of passages to Powell Cave and mapped more than 2 km of passages in Neel Cave. (Walsh and Kastning, 1977; anonymous, 1981; Hooper, 1981, 1982; Veni, 1981a,c,d; Atkinson, 1982; Waters, 1982).

Few geologic studies had been carried out at the Powell Cave System. Frank (1962a, 1963, 1965b) and Frank and Frank (1963) collected sediments and fossils of vertebrates from Powell Cave, but analysis of these deposits yielded little paleoclimatological information except for an indication that climates became progressively drier during the Pleistocene. Palmer (1975) reviewed the map of Powell Cave as part of his study on the origin of maze-like caves, and he identified sections of maze in the cave as examples of the "network" variety of maze. Although his study is based on extensive fieldwork, Palmer did not visit Powell Cave in particular.

DEVELOPMENT OF CAVES

Descriptions of Caves

Powell Cave

Powell Cave and Neel Cave, which were recently linked by exploration, together total more than 19.5 km of mapped passages (Plate 6). The maximum known extent of the entire cave is greater than 3.2 km, ranking second in this regard among the long caves of Texas (Appendix C). That part of the system known as Powell Cave has a mapped length in excess of 17.5 km. The single, natural entrance to this section is a 6 m-deep pit at the bottom of which is a 10 m-long crawlway connecting with one of the upper-level mazes in the cave. A 14 m-deep shaft intersects the Main Stream Passage 73 m to the southeast of the entrance and was supposedly dug by treasure-seekers in quest of the lost San Saba or Bowie Silver Mine (Dobie, 1924, 1930).

The cave consists of three components differing greatly in morphology and developed along distinct, stratigraphic horizons. These include (1) the Main Stream Passage, (2) the Crevice Passage, and (3) several of the sections of mazes (Plate 6).

The Main Stream Passage lies along the southwest flank of the cave. It can be entered from the mine shaft, at the southeast end of the Crevice Passage, or by a short side passage connecting to the southernmost maze. This passage has been mapped more than 1.6 km to the northwest (upstream) from the Crevice Passage to the present limit of exploration, where it continues and presumably connects with Silver Mine Cave. Approximately 440 m downstream from the Crevice Passage, the Main Stream Passage enters the siphon through which the connection between Powell and Neel Caves was made in 1982. Previously, it had been shown by tracing of dyes that water from this point flows southeast through Neel Cave to Lakeside Spring (Russell and Reddell, 1968). Of the few side-passages found along the Main Stream Passage, most reconnect, forming short loops and bypasses. In general, the Main Stream Passage averages 4 to 8 m wide and 1.5 to 2.5 m high throughout its length. However, the last 120 m of passage just upstream of the siphon is much reduced in cross-section, with average dimensions of 1.5 m wide by 1.5 m high.

The Main Stream Passage has been nearly severed by a zone of collapse 60 m southeast of the mine shaft. This area lies beneath an elliptical doline 35 m by 46 m in size and 4 m deep. A crawlway of small dimensions and containing the stream passes through breakdown beneath this collapse. The Main Stream

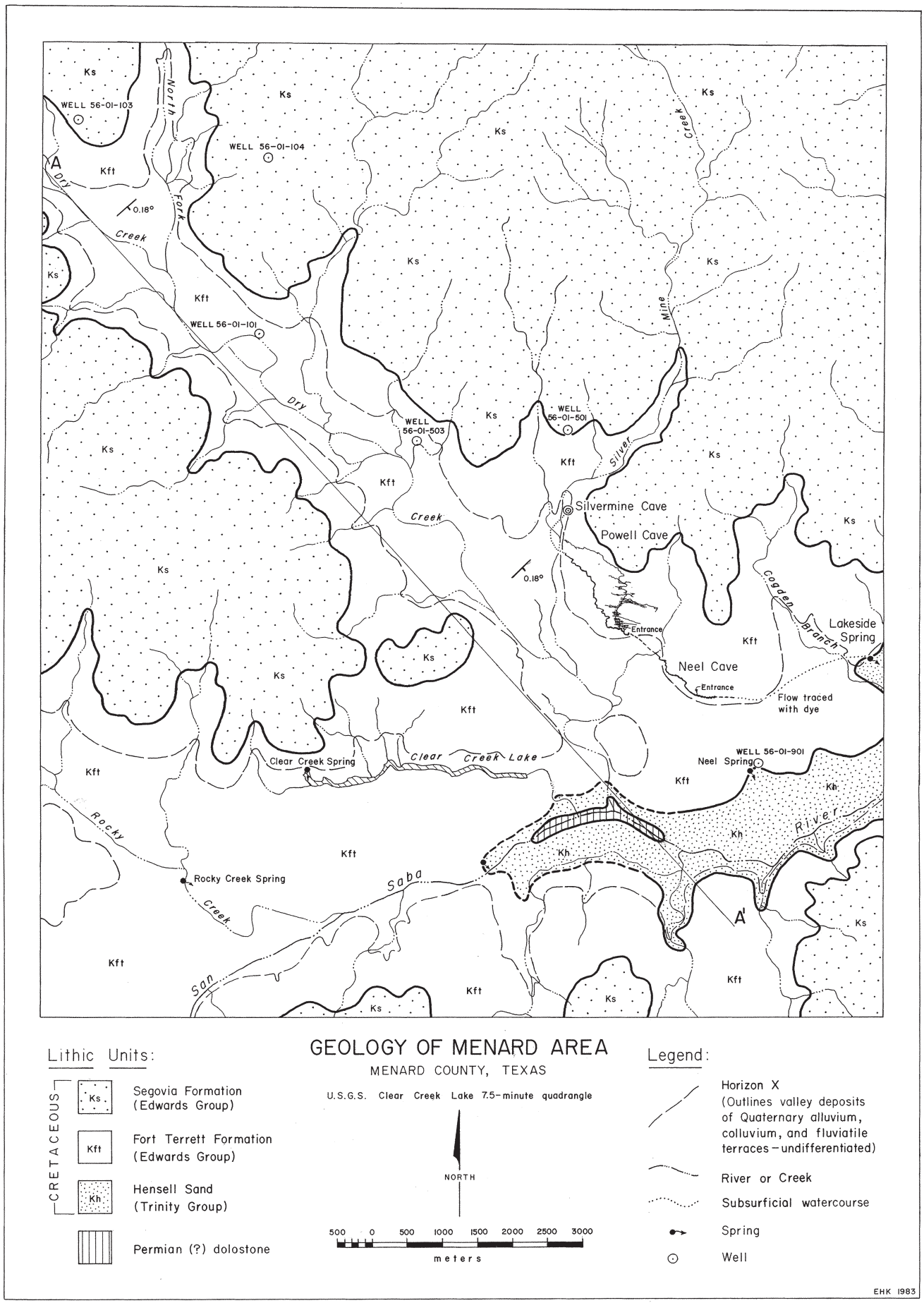


Figure 45: Geologic map of the Menard area. Caves of the Powell Cave System and major springs are indicated. Geology mapped from 1:20,000 aerial photographs in Coffee and others (1967).

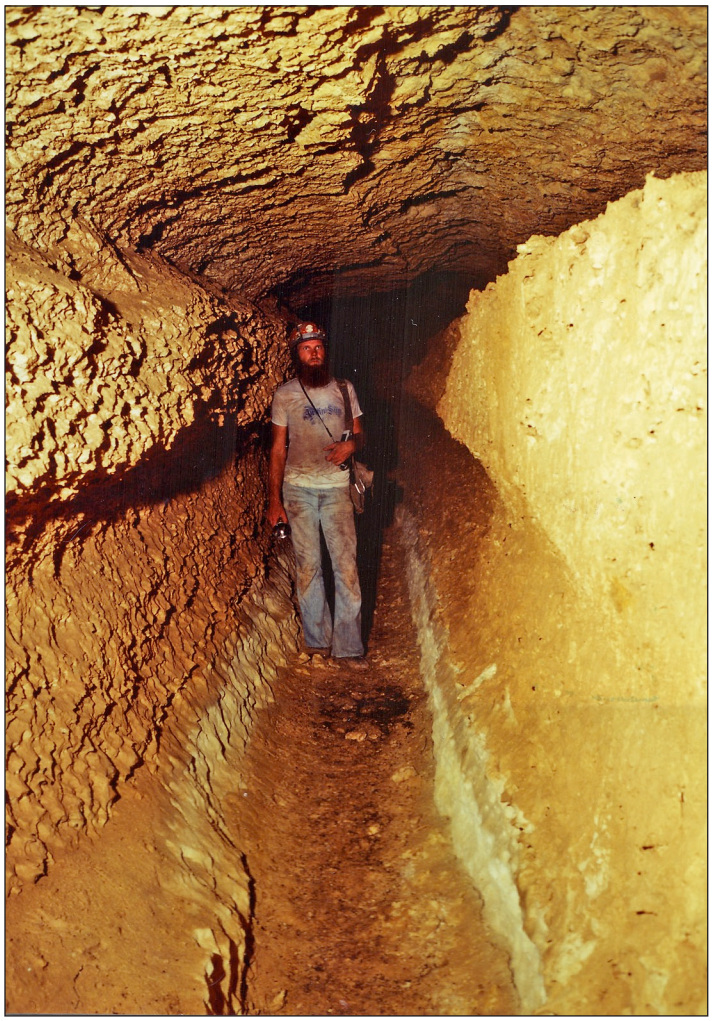


Figure 46: Crevice Passage in Powell Cave. A joint that has guided the orientation of the passage is visible in the ceiling and along the right wall. The conduit which initially developed in the phreatic zone is visible in upper part of passage. The incised channel beneath this has been cut by a vadose stream. Pitted walls and coating of clay and silt have resulted from periodic flooding.

Figure 47: The Crevice Passage (foreground) and overlying passages of the maze near the Big Bend in Powell Cave. Mazes developed within the most soluble and porous beds or along bedding-plane partings and are strongly guided by joints. A thin mantle of silt and clay deposited by floodwater occurs in most passages.



Passage contains banks of sediment alongside the stream that meanders within the conduit. Normal flow of the stream is estimated at 1 liter per second.

A second, major passage in Powell Cave is a sinuous, narrow conduit known as the Crevice Passage (Plate 6; Figures 46, 47, and 48). At its southeast end it connects with the Main Stream Passage in the vicinity of the zone of collapse. The Crevice Passage, with a total length of 2.2 km, trends north for a distance of 670 m and then continues generally N 65° W for an additional 840 m to a series of low crawlways. The width of this passage is approximately 1 to 2 m throughout its length; however, its height gradually decreases along its length from 4 m near the entrance to the cave to less than 1 m at its northwest terminus. Decrease in height of the passage is due to a rise in elevation of its floor, as the elevation of the ceiling remains relatively horizontal. The Crevice Passage is the common "thread" linking together all sections of mazes in the cave (Plate 6; Figures 47 and 48) and connecting the mazes to the Main Stream Passage. The Crevice Passage is dry and no running water has ever been observed in it.

The remaining 13.2 km of passages in Powell Cave consist of mazes. These have developed along the uppermost levels in the cave and branch east and west from the Crevice Passage (Plate 6). All known mazes are located northeast of the Main Stream Passage, and most have developed along a single level. However, in some areas, such as north of the Metro, other levels of mazes occur above the former level. Generally, passages in the mazes average 2 to 4 m in width and 0.2 to 1.5 m in height (Figures 47, 48, and 49).

The Crevice Passage and mazes contain thin deposits of sediment and only a small amount of breakdown. In contrast to this, breakdown blocks are locally present along the entire length of the Main Stream Passage. There are only minor amounts of speleothemic deposits throughout the cave.

Powell Cave is confined to a 12 to 15 m-thick interval of the upper Fort Terrett Formation, and nowhere does the cave lie more than 20 m below the surface. As a consequence, roots of mesquite have penetrated the ceilings of passages at all levels.

Neel Cave

Neel Cave, as a section of the Powell Cave System, has a mapped length of more than 2 km. It consists of a single conduit containing the stream which leaves Powell Cave (Plate 6). The only entrance to Neel Cave is a 13 m-deep pit which leads directly to the level of the stream. From this point the passage entends northwest (upstream) for 1130 m to a series of siphons which were penetrated by diving in 1982, leading explorers into Powell Cave. The first siphon is 460 m to the southeast of the downstream siphon in Powell Cave, and the passage between the siphons is estimated to be at least 600 m long. Downstream from the entrance, Neel Cave has been explored and partially mapped for at least 2 km along a single conduit. Water flowing along this passage resurges at Lakeside Spring (Figure 45).

Neel Cave is morphologically similar to the Main Stream Passage in Powell Cave, having average cross-sectional dimensions of 4 to 8 m in width and 1.5 to 2.5 m in height. Only a few side-passages have been found and these either end abruptly or reconnect with the main passage. The discharge of the stream coursing through Powell and Neel Caves is nearly constant along its length. Roots of mesquite penetrate into Neel Cave throughout its length.

Silver Mine Cave

Little has been known about Silver Mine Cave owing to past difficulties in acquiring permission to visit the cave. However, the cave is presently being explored and mapped by speleologists. It is reported to have a long passage with a stream which is along trend with the Main Stream Passage of Powell Cave and presumably is the upstream continuation of this water course. Entrance to Silver Mine Cave is through a mine-shaft near Silver Mine Creek (Figure 45). At least 600 m of passage have been explored to date. A nearby doline, known as the Egg Shaped Sink, reportedly behaves as a swallet during flooding along Silver Mine Creek. It is said to be a former entrance to Silver Mine Cave (Lindsley, 1967b).

Meteor Hole

The Meteor Hole is a filled, collapsed doline that opened on 30 August 1938 in response to a month of heavy rains (Blakemore, 1939). It was a circular opening 8 m in diameter and 9 m deep with walls composed of a bedded, cobble-rich conglomerate with an impure, limy matrix. Unconfirmed reports indicate that some passages leading from this doline were explored prior to slumping and filling (Russell and Reddell, 1968).

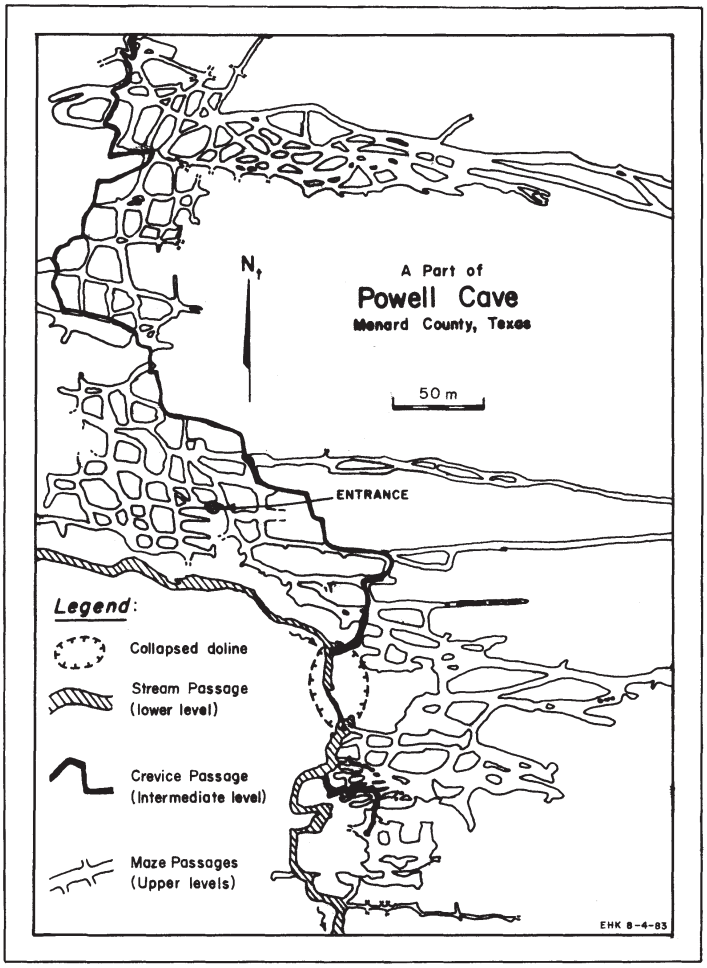


Figure 48: Map of part of Powell Cave in the vicinity of the entrance showing the Stream Passage (lower level), Crevice Passage (intermediate level), and maze-like passages (upper levels). See also Plate 6.

Lakeside Spring

Discharge from Lakeside Spring is comparable to that of the stream in Powell and Neel Caves. The spring issues from an opening 2 m wide and at least 1 m high. An artificial dam built to create the pool at the spring presently keeps the water level at the top of the opening, barring exploration of the passage.

Topographic Relationships

The Powell Cave System lies beneath a relatively level, erosional surface within the valleys of Dry Creek, Cogden Branch, and the San Saba River (Figure 45). Local baselevel is along the San Saba River, approximately 21 m below the level of the resurgence at Lakeside Spring. The southeast-trending valley of Dry Creek is remarkably linear, and it is parallel to and nearly at grade with the regional dip (Figure 50). Dry Creek turns abruptly to the south at an apparent elbow of capture approximately 1.5 km north of Beyer Crossing. Nearby, Clear Creek changes from an eastward course to a southward course in a similar fashion (Figure 45).

The gradient of the San Saba River (0.11°), as it flows northeast through the area, is slightly less than the regional dip (0.18°) in the vicinity of the Powell Cave System. At Beyer Crossing the river has eroded through the entire Fort Terrett Formation and has exposed Permian rocks along a short reach (Figure 45). Downstream from Beyer Crossing the river flows nearly down the dip for 2.4 km (Figure 50) and then turns northeast as before.

Speleogenesis

Configurations of passages in the Powell Cave System (Plate 6) suggest strong, lithologic and structural control. Many passages, particularly in the mazes, are linear in plan-view. In addition, differences in morphology among sections of the cave suggest distinct stages during development of the cave.

Lithologic Control

The mazes on upper levels and the Main Stream Passage have developed within distinct beds of the Fort Terrett Formation and are generally conformable with them. Most passages in the mazes occur within a 0.7 to 1.5 m-thick, highly burrowed unit (Figure 49) or along bedding-plane partings within porcelaneous and dolomitic beds of limestone (Figure 47). The Main Stream Passage has developed along a burrowed zone of thin-bedded to massive limestone.

The Main Stream Passage in Powell Cave and the major conduit of Neel Cave course directly down the dip of the strata. The stratigraphic horizon of this passage is the same as that of major springs of the area (Figure 50). The passages containing the streams and the springs are 5 to 10 m above the San Saba River, the local baselevel. Development of the passages containing streams along the dip and the absence of known conduits between this horizon and baselevel indicate that the passages are perched. Conceivably a massive, dolomitic bed or sequence of dolomitic and marly beds just beneath the passages has prevented complete adjustment of developing conduits to entrenchment of the San Saba River and its tributaries. Support for this is provided by data from wells in the area which show that wells generally yield water from the upper Fort Terrett Formation of from the underlying Paleozoic rocks, with few wells yielding water from the lower beds of the Fort Terrett (Baker and others, 1965).

Structural Control

The passages containing streams in Powell and Neel Caves meander slightly because they alternately follow fractures of two primary sets. One set is oriented at S 70° E to S 85° E, and the other at N 5° E to N 20° W. Joints that have guided development of passages are visible in the ceilings along the length of the conduits. In some places, such as downstream from the Neel Cave entrance, breakdown has been dislodged along zones of fractures of the S 70° to 85° E set, indicating that local zones of structural weakness and increased permeability occur along this set of fractures. Another example of structural weakness within a fracture zone is the area of collapse in the southernmost mazes of Powell Cave and along the adjacent stream-passage on the lowest level.

Control by fractures is most obvious in the mazes. Here passages are oriented along the two aforementioned sets of fractures. Fractures of both sets have nearly identical frequency and spacing; however, fractures of the easterly orientation are more extensive and have resulted in longer segments of passages in those directions.

Nearly all segments of the Crevice Passage follow fractures of the two sets. In many places segments of the Crevice Passage and overlying passages of the mazes are guided by the same fractures (Figures 46, 47, 48; Plate 6).

Figure 49: Typical passage of the maze in the vicinity of the entrance to Powell Cave. Note highly burrowed and porous character of rock at walls and ceiling and clastic sediment on floor.



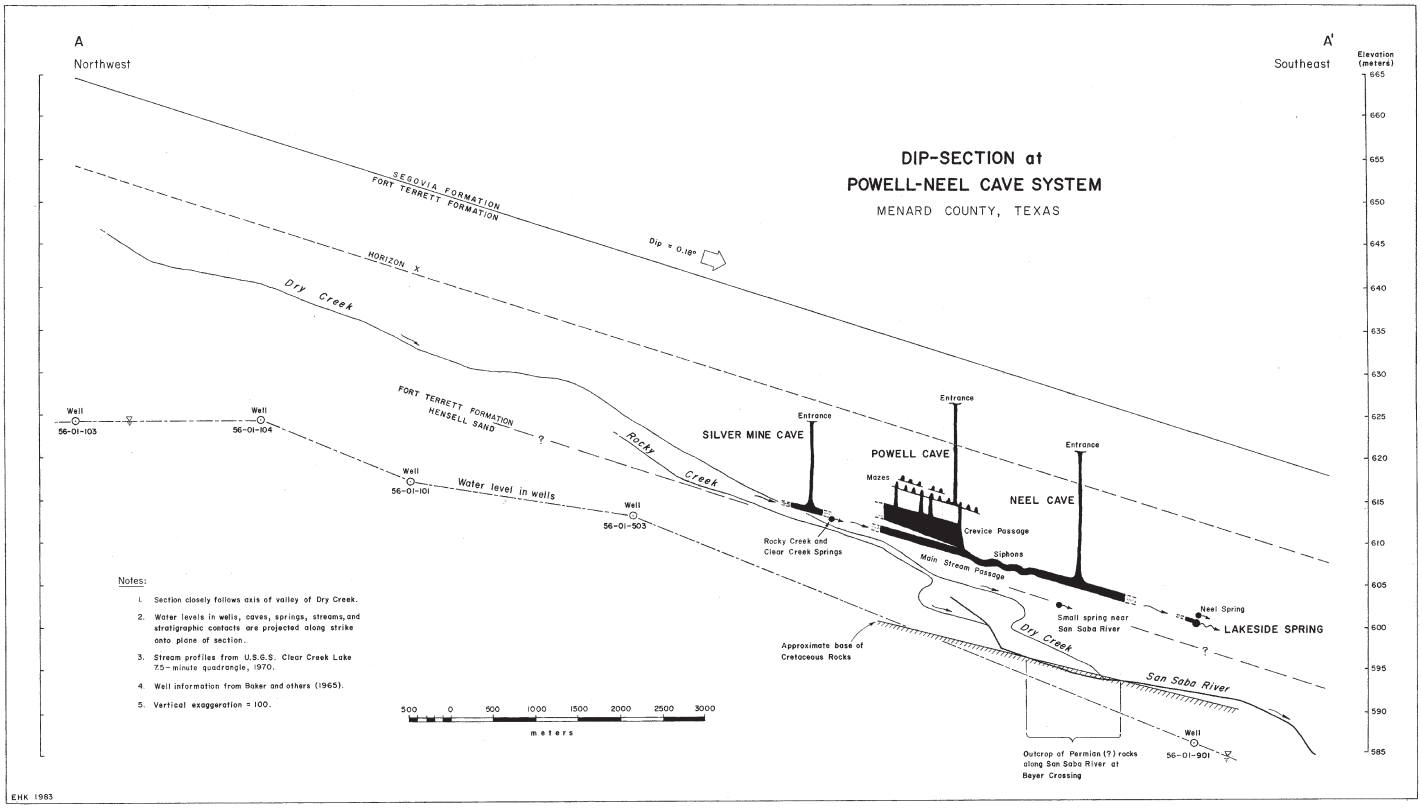


Figure 50: Dip-section long axis of valley of Dry Creek. Profile of creeks, contacts between lithic units, and horizons of caves and springs, and piezometric surface are projected onto the plane of section.

An attempt at mapping fractures from aerial photographs provided little information in the Menard area. Fractures apparently have weak expression as photolineaments. A few valleys and scarps on hillsides are aligned along fractures of the two sets. The most prominent alignment is that of Clear Creek Lake along the easterly trend (Figure 45).

Fractures appear to be caused by tensional stresses associated with regional warping of the area. Although the tectonic history of the fractures remains undetermined, they may be upward extensions of pre-existing fractures which may have originated in the underlying Paleozoic rocks during the Llano Uplift; or they may be associated with regional uplift of central Texas during the Cretaceous or Tertiary Periods.

The dominant, structural control acting on the development of caves has been the regional dip (0.18° in a S 42° E direction). The average trend of the mapped passages containing streams in Powell and Neel Caves is exactly aligned with the direction of dip. To the east of the Neel Cave entrance, however, the master conduit begins to trend east and northeast toward the resurgence at Lakeside Spring. This suggests that other factors, such as the evolution of the San Saba valley, may have influenced the orientation of the passages in the eastern part of the system.

Hydrogeologic Control

The Powell Cave System drains a large area of the basin of Dry Creek. The exact extent of this drainage can not be ascertained until the source of the stream in Powell and Neel Caves is precisely determined. However, the large flowrate (1 liter/sec) of the perennial stream in the cave suggests that flow through the master conduit may represent water lost by infiltration within the valleys of Dry Creek and its tributary, Silver Mine Creek.

Nearly conformal development of Dry Creek on strata of the upper Fort Terrett Formation suggests that infiltration contributing to the stream in the cave may take place well up the valley of the creek, perhaps 10 km or more beyond the explored, upstream limits of the cave-system. The collapse of the Meteor Hole may have been related to such a system of drainage.

The minimum extent of the main stream from its explored position upstream in Powell Cave to Lakeside Spring is 6 km. This is certainly the longest known stream in a cave in Texas and it is long even by worldwide standards. This length indicates that drainage in the subsurface is well integrated and that the Powell Cave System is in a late stage of development.

The Crevice Passage of Powell Cave was once a tributary to the Main Stream Passage. Moreover, it was once a single conduit characteristic of those developed in nearly horizontal strata in the phreatic zone. The uppermost part of the Crevice Passage is elliptical in cross-section and developed along intersections of bedding-plane partings and vertical joints (Figures 46 and 47). As the piezometric surface dropped in response to lowering of local baselevel, the Crevice Passage was drained of phreatic waters, and flow continued through the passage as a free-surface, vadose stream. Entrenchment of the stream, in response to a difference in elevation between the Crevice and Main Stream Passages, has excavated a channel into the floor of the original, phreatic conduit that eventually became graded toward the Main Stream Passage (Figures 46 and 47). Any flow in the Crevice Passage has since been pirated to the Main Stream Passage, presumably at a point upstream (northwest) of the explored limits of both passages.

Development of Mazes

Powell Cave has been significantly modified by periodic flooding. The most obvious manifestations of this are the extensive mazes branching from the upper level of the Crevice Passage.

A *maze-pattern* "consists of a labyrinth of intersecting passages of rather uniform character that form closed loops" (Palmer, 1975, p. 57). This is in contrast to a *branchwork-pattern*, where conduits intersect or converge as tributaries along the direction of flow. A true maze-cave results from simultaneous, rather than sequential, enlargement of fractures and bedding-plane partings (Palmer, 1975).

Mazes in Powell Cave are guided by both joints and bedding-plane partings. *Network-mazes*, consisting of angular grids of intersecting fissures (Palmer, 1975), are most common where joints have been solutionally widened. *Anastomotic mazes*, characterized by curvilinear tubes intersecting in a braided configuration (Palmer, 1975), are most common where bedding-plane partings have been solutionally enlarged.

The mazes in Powell Cave have apparently developed in response to flooding upstream from a zone where the Main Stream Passage had been partially blocked by collapse. Upon enlargement of the Main Stream Passage to critical dimensions, collapse of ceilings occurred along a north-trending set of fractures just downstream from where the Crevice Passage meets the Main Stream Passage (Plate 6). This collapse extended to the surface, forming the large doline there. Collapse caused a significant constriction in the Main Stream Passage so that the conduit could no

longer transmit the high discharge incurred during periods of heavy precipitation and runoff. Water periodically became impounded behind the constriction. Flooding was sufficient to force water to rise and fill the Crevice Passage. Because floodwater is chemically and hydrodynamically aggressive (Palmer, 1971), it can enlarge pre-existing fractures and partings at rates greater than those for water moving slowly under laminar conditions in the phreatic zone. Passages of mazes in Powell Cave have been rapidly excavated along strata which were lithologically most suitable to dissolution,such as highly burrowed units, and along bedding-plane partings of high permeability. These passages have been guided horizontally by available north- and east-trending fractures.

Several easily observed features in passages of the mazes provide evidence for development by floodwater. Nearly all passages terminate abruptly in bedrock and are not segments of former paths of flow now truncated by clastic fill or collapse. Development of mazes decreases northward, away from the point of collapse in the Main Stream Passage, indicating that, as floodwater ponded further up the Crevice Passage, it became more saturated with respect to calcite and perhaps more tranquil, thereby losing its power to dissolve bedrock. Walls of the Crevice Passage and passages in mazes have thin coatings of clay and silt (Figure 46), and floors of most passages are mantled by a thin (1-5 cm) deposit of this material (Figures 47 and 49).

A section of maze extends along the east side of the major zone of collapse and connects with the Main Stream Passage south of the collapse through a short passage similar to the Crevice Passage in cross-section and stratigraphic position. Apparently this section of maze developed as a route of diversion through which floodwater bypassed the constricted area.

Chronology of Speleogenesis

The following chronological sequence for the origin of the Powell Cave System is postulated, based on the foregoing observations and interpretations:

1. Erosion of the Edwards Plateau was accelerated at the time of Balcones Faulting (Miocene). Streams of the Colorado River basin, including the San Saba River and its tributaries, began to cut into the surface of the plateau, removing the Late Cretaceous rocks and part of the Segovia Formation. The Fort Terrett Formation was exposed in the valleys of Dry and Rocky Creeks.

2. Dry and Rocky Creeks developed down the dip-slope on the Fort Terrett Formation. Their valleys are remarkably linear in the direction of dip.

3. Infiltration of meteoric water into the Fort Terrett Formation within the valleys of these creeks led to integration of flowpaths of groundwater and enlargement of north- and east-trending fractures. A single, master conduit developed in the phreatic zone along alternating sets of fractures. Despite its angularity the overall trend of the conduit remained parallel to the dip and comformable to the bedding. Tributary-passages such as the Crevice Passage in Powell Cave and perhaps passages in Silver Mine Cave were also established at this time. The conduit containing the main stream in Neel Cave deviated from the direction of dip to flow eastwardly and northeastwardly to Lakeside Spring. Apparently this change in course was due to the former position of the San Saba River northwest of its present location in the reach between Beyer Crossing and Lakeside Spring. By the time the San Saba migrated southeastwardly, the master conduit had become well established.

4. As the San Saba River cut deeper into the Fort Terrett Formation, waterlevels in the cave dropped and thereafter the Main Stream Passage and Crevice Passage were in the vadose zone. Incision by a vadose stream lowered the floor of the Crevice Passage, grading it toward the Main Stream Passage at a lower elevation. Piracy beheaded the former stream in the Crevice Passage, channeling its flow directly to the main stream at a point upstream of presently explored passages.

5. Enlargement of the Main Stream Passage in Powell Cave and its counterpart in Neel Cave led to collapse of ceilings in densely fractured zones. One such collapse, extending from the Main Stream Passage to the surface nearly blocked the flow of groundwater through the master conduit. As a consequence, sections of the cave upstream of this collapse became subjected to periodic flooding during heavy rain and increased runoff.

6. Backflooding and ponding along the Crevice Passage produced extensive anastomotic and network-like mazes in the upper levels of the cave. These developed along highly soluble beds and fractures.Eventually floodwaters were diverted around the zone of collapse through passages in the mazes.

SUMMARY

Morphologically distinct sections of the Powell Cave System represent three stages of speleogenesis: (1) enlargement and integration of the master conduit and its tributaries in the phreatic zone, (2) lowering of the floor of the Crevice Passage by a free-surface stream under subsequent vadose conditions, and (3) excavation of upper-level mazes by periodic flooding. In Powell Cave the uppermost passages are the youngest, a situation reversed from that in many other caves where horizons of passages are correlative with pauses in the lowering of baselevels by surficial streams.

8
BEND AREA

DESCRIPTION OF AREA

The Bend area and Burnet area (Chapter 9) are two of the most significant localities for karst within the Paleozoic, carbonate terrane of the Llano Basin. The Bend area, along the Colorado River in southeastern San Saba County and southwestern Lampasas County (Plate 1), is bounded on the south and north along lines of 31°1'35" and 31°4'15" north latitude respectively, and on the east and west along lines of 98°27'43" and 98°30'00" west longitude respectively (Figure 51). The area lies within the Gorman Falls 7.5-minute U.S.G.S. quadrangle and is drained by tributaries of the Colorado River, including Tie Slide, Gorman, and Spicewood Creeks, and several short draws flowing directly into the Colorado River (Figure 51).

Geologic maps that include the Bend area have been published at scales of 1:10,000 to 1:333,474 (Plummer, 1940; Cloud and Barnes, 1946a, plate 9; Dietrich and Lonsdale, 1958, plate 1; Barnes, 1976a). Exposed rocks in the Bend area primarily consist of beds of the Gorman and Honeycut Formations of the Ellenburger Group (Ordovician). The Chappel Limestone and Barnett Formation (Mississippian) crop out in the extreme northwestern corner of the area, and travertine deposits and alluvial deposits associated with terraces, both Quaternary in age, lie along the Colorado River (Figure 51).

Geologic sections from within the Bend area are described in Cloud and Barnes (1946a, p. 231-238) and Hendricks (1952, p. 23-30). The general stratigraphic relations among the facies and formations of the Ellenburger Group are given in Figure 8 and Appendix B. Several studies are available on the Ellenburger units outlining (1) additional stratigraphic information (Dake and Bridge, 1932; Sellards, 1932a; Barnes and Cloud, 1945; Bridge and Cloud, 1947; Barnes, 1953), (2) depositional history (Cloud and Barnes, 1946b, 1948, 1957; Barnes and Cloud, 1972), (3) subsurface correlations (Hendricks, 1940, 1942, 1964; Bradfield, 1964), and (4) physical, chemical, and petrographic properties (Udden and Waite, 1927; Parmelee, 1946; Goldich and Parmelee, 1947; Barnes and others, 1959).

The vicinity of Gorman Falls is the type-locality of the Gorman Formation, consisting of thick- to thin-bedded, cherty limestone and dolostone which contain thin, arenaceous beds. The Gorman Formation is subdivided into a lower, dolomitic facies and an upper, calcitic facies (Cloud and others, 1945; Cloud and Barnes, 1946a). The dolomitic facies in the Bend area is 25-66 m thick, but only the upper few meters are exposed along the walls of the valley of the Colorado River (Figure 51). This unit consists of microgranular to finely crystalline dolostone ranging in color from shades of pink and gray to yellowish gray. It contains few fossils and weathers to smooth, crudely sphenoidal blocks or to hackly or blocky ledges (Cloud and Barnes, 1946a; Barnes, 1976a). The overlying calcitic facies, which is exposed over most of the uplands of the Bend area (Figure 51), is 76-117 m thick, and consists of upper and lower units of aphanitic, thick- to thin-bedded, light gray limestone and a middle unit of microgranular to finely crystalline dolostone (Figure 8). The uppermost bed of the calcitic facies is a massive limestone (Cloud and Barnes, 1946a; Barnes, 1976a). A zone of chert containing fossils of the sponge. *Archaeoscyphia* forms a useful stratigraphic marker approximately midway in the Gorman Formation and about 50 m above the contact between the dolomitic and calcitic facies (Figures 8 and 51).

The overlying Honeycut Formation, exposed in the northwestern and northeastern parts of the Bend area (Figure 51), consists of thick- to thin- bedded, cherty limestone and dolostone. The basal part is an interbedded sequence of limestone and dolostone, whereas the upper part is largely microgranular to finely crystalline, light gray to yellowish gray dolostone alternating with thin beds of limestone (Figure 8). Fossils are common and are preserved in nodules of chert (Cloud and Barnes, 1946a; Barnes, 1976a).

A massive deposit of Quaternary travertine is found at Gorman Falls where Gorman Creek plunges approximately 15 m to the Colorado River (Plummer, 1943, frontispiece). The deposit forms a broad, elevated terrace, 350 m long and 150 m wide, which has filled the valley of the creek to a depth of about 20 m at the confluence of the creek with the Colorado River (Cloud and Barnes, 1946a, plate 35B, p. 442; see also Figure 51). The source of the dissolved carbonate precipitated at the falls is Gorman Spring, 790 m to the southwest along Gorman Creek (Figure 51).

The dip of beds of the Ellenburger Group in San Saba and Lampasas Counties is generally 1.0° to 1.4° to the northeast (Levorsen, 1941; Sellards and Hendricks, 1946; Turner, 1957, figure 8, p. 74). However, locally within the Bend area, the dip varies between 0° and 4° to the northwest (Cheney, 1929, plate 7, 1940, figure 9, p. 108; Cloud and Barnes, 1946a, plate 9; see also Figure 51).

The Ellenburger rocks of the Bend area are cut by a series of *en echelon*, normal faults ranging in dip from 60 to 90 degrees and striking north-northeast (Figure 51). These are part of the Llano system of faults formed during the Pennsylvanian Period in response to tensional epeirogenic stresses (Powers, 1928; Sellards, 1934, p. 85; Cloud and Barnes, 1946a, p. 120-121; Eardley, 1951, p. 228; Cheney and Goss, 1952; Orr, 1962; see also Chapter 2 for summary of Paleozoic tectonism). As a result, much of the outcrop of the Ellenburger in the region has been compartmentalized as horsts and grabens.

The Ellenburger units are well-jointed with joints spaced a few meters apart on the average and dipping between 70° and 90°. Folds in the area are few and those that exist are related to the development of collapsed dolines during the Late Ordovician to Devonian (Cloud and Barnes, 1946a).

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

The caves of San Saba County have been catalogued by Reddell and Estes (1962) and Reddell (1973) who list 51 caves in the Bend area, one of the most cavernous localities in the county and in the state (Reddell, 1973, p. 5). A compilation of caves in Lampasas County has not yet been published. Additional exploration has raised the total number of known caves in the Bend area to at least 76, but only eight of these exceed 50 m in length.

Gorman Cave is one of the best known and most frequently visited caves in Texas (Suhrstedt, 1979), and early explorations of it apparently date to the 1860's and 1870's (Reddell, 1973, p. 56). One of the earliest reports describing the exploration of caves in Texas was written by Cummins (1890) who briefly described Gorman Cave (quoted in Reddell, 1973, p. 56). The cave was visited and discussed often during the formative years of organized speleological activity in Texas (anonymous, 1934; Cloud and Barnes, 1946a; Riggs, 1951; Balcones Grotto, 1955), and descriptions and maps have appeared in several guides to the caves of Texas (Reddell, 1964b; Smith, 1971a; Fieseler and others, 1978).

Many additional caves were found and recorded as speleological activity increased in the area (Estes, 1962, 1968; Jasek, 1969, 1975a, 1977b; Kastning, 1974). Estes (1962) and Russell (1968) have commented on the unusual diversity of caves located here, and Beck (1970) note the dynamic characteristics of the Colorado River in the vicinity of caves near baselevel. The influences of lithostratigraphy and tectonism on speleogenesis in the Bend area were examined by Kastning (1981b, 1983c, 1984).

Massive deposits of travertine are common in and near the Bend area. Large accumulations occur at Gorman Falls (Figure 51; see also Plummer, 1943, frontispiece for photograph; Suhrstedt, 1979), and at Post Oak Falls on the Colorado River, 2.4 km south-southeast of the Bend area, and elsewhere (Paige, 1912a, plate 7). Plummer (1939, 1944b, 1945) and Barkley and Plummer (1944) briefly discuss the origin of deposits of travertine in central Texas and Gale (1912, 1932), Mansfield and Boardman (1932), and White (1948a) describe small, nitrate-bearing caves within deposits of travertine at Post Oak Falls and elsewhere in San Saba County.

Areas of collapse into underlying solutional voids are common in Paleozoic rocks of the Bend area and neighboring localities in the Llano region. Plummer (1943) has described and mapped four collapsed dolines within the Ellenburger beds that average 50 to 80 meters in diameter and contain deposits of

Mississippian Chappel Limestone and Barnett Formation. The dolines have been termed "structural sinks" by Cloud and Barnes (1946a) because Mississippian rocks dip into the dolines as a result of contemporaneous collapse of these beds and those of the underlying Ellenburger. Three such collapsed dolines in the Bend area (Figure 51) have been mapped and described (Cloud and Barnes, 1946a). Further remarks on these and other paleokarstic features in central Texas may be found in Quinlan (1978).

Hydrogeologic investigations of the carbonate rocks in the Bend area have been minimal. Aside from inventories of wells (George and Shafer, 1939), brief overviews of the Ellenburger aquifer (Plummer, 1946; Mount and others, 1967), and an inventory of major springs (Brune, 1975), little is known about the dynamics of groundwater in the region.

DEVELOPMENT OF CAVES

Forty-two caves were investigated in the area during this study (Table 10). They are grouped according to morphology and location as follows:

(1) Single caves in the southeast—includes three generally long caves (Gorman, Lemons Ranch, and Clarks Branch Well Caves) and three small, but associated, caves on the uplands (El Sotanito Grande, Yates Goat and Barnes Bat Caves).

(2) Gorman Fissure System —four small, fissure-like caves on the upland near Gorman Cave.

(3) Bluff Caves—five horizontal caves opening onto the bluffs along the Colorado River.

(4) Northwest caves—six caves on the hillsides northwest of Gorman Falls.

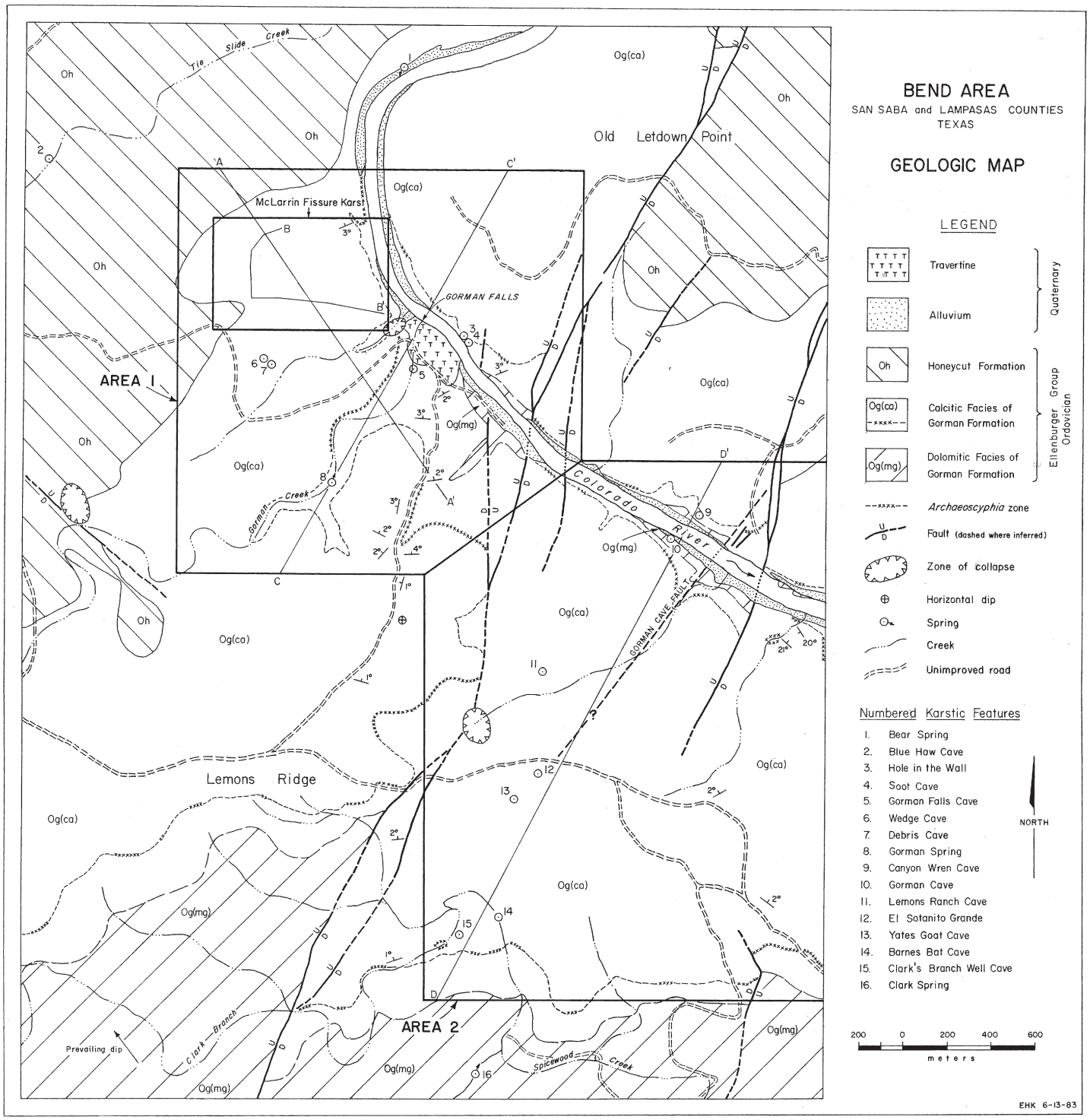


Figure 51: Geologic map of the Bend area, San Saba and Lampasas Counties. Detailed maps of Areas 1 and 2 are shown in Figures 69 and 64 respectively. Cross-sections A-A', B-B', C-C', and D-D' are shown in Figures 89, 74, 90, and 91 respectively. Geology modified from map of Cloud and Barnes (1946a, Plate 9).

Table 10: Caves of the Bend Area

Cave	Cave Type[a]	Length (m)	Depth (m)	Map figure	Rose diagram	Reference[b]
Gorman Cave	C	914	10	52, 64	93A	1: 52. 54-58
Lemons Ranch Cave	C	168	30	59	93B	1: 78-81
El Sotanito Grande	P	69	32	60	93C	1: 101, 102
Yates Goat Cave	P	12	20	61	93C	1: 119, 120
Barnes Bat Cave	C	158	10	62	93C	1: 15, 16
Clarks Branch Well Cave	C	335	12	63	93D	1: 29-31
Caves of the Gorman Fissure System				64, 65	93E	
Crevice Cave	F	5	12+	65	93E	1: 35
McLarrin Flowstone Fissure	F	5	17	65	93E	1: 83
Toad Frog Falling Floor Fissure	F	5	5+	65	93E	1: 108
Coon Cave	F	5	5+	65	93E	1: 32
Bluff-Caves					93F	
Canyon Wren Cave	B	20	2	64, 67	93F	
Soot Cave	B	64	4	70	93F	2: 127-129
Hole in the Wall	B	11	3	71	93F	2: 127-129
Skull Cave	B	14	5	72, 73	93F	1: 100
Climbup Cavelet	B	5	2	72, 73	93F	
Northwest Caves					93F	
Copperhead Cave	R	18	8	73, 74, 75	93F	1: 32-34
El Sotano Lodoso	P	33	28	73, 74, 76	93F	1: 101-103
Gorman Sink	R	7	6	73, 74, 77	93F	1: 59, 60
Debris Cave	R	14	1	78	93F	
Wedge Cave	R	16	6	79	93F	1: 113
Blue Haw Cave	B	12	5	80	93F	
Gorman Falls Cave	C	82	27	69, 81	93G	1: 59, 6
Caves in Travertine						
Goromlette Waterfall Cave	T	23		82		1: 59. 62-63
Mini Cave	T	8				3: 49. 53-54
Caves of the McLarrin Fissure Karst				73, 74	93H	
X Fissure	F	9		85A	93H	
Walk By Pit	F	1		85A	93H	
Meander Tube Fissure	F	20		85B	93H	
Offset Fissure	F	13		85C	93H	
Honeycombless Fissure	F	8		85D	93H	
Two Part Fissure	F	10		85E	93H	
Cobweb Fissure	F	9		85F	93H	
Fern Fissure	F	9		85C	93H	
King Fissure	F	24		85H	93H	
Slumped Rock Fissure	F	7		85I	93H	
Pokeweed Fissure	F	18		85J	93H	
Three Entrance Fissure	F	11		85K	93H	
Dagger Cave	F	12		85L	93H	1: 37-38
Gourd Fissure	F	10		85M	93H	
Covered Crawlway Fissure	F	10		85N	93H	
Snail Fissure	F	16		85O	93H	
Blacktail Pit	F	10		85P	93H	
Crack Pot	F	18		85Q	93H	

[a]Type of cave (for description of type, see discussion on speleogenesis):
C= cave with master conduits R= cave with a single room
P= pit-cave F= fissure-like cave (crevice)
B= bluff-cave T= cave in travertine

[b]Most recent and complete reference to cave with page numbers:
1. Reddell (1973)
2. Kastning (1974)
3. Jasek (1969)

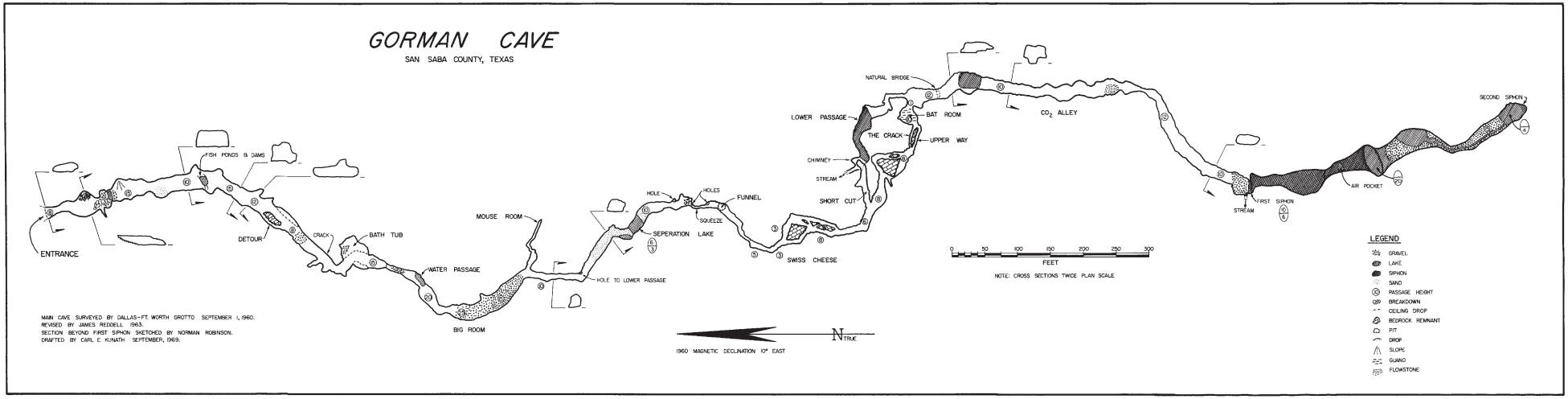


Figure 52: Map of Gorman Cave. Reddell (1973, p. 56A).



Figure 53: Main entrance to Gorman Cave. Note massive bed of dolostone at left-center showing sphenoidal fracturing.

(5) Gorman Falls Cave—a cave lying beneath the valley of Gorman Creek near Gorman Falls.

(6) Travertine caves—two small caves developed within deposits of travertine at Gorman Falls.

(7) McLarrin Fissure System—eighteen small fissure-like caves in the uplands immediately west-northwest of Gorman Falls

Descriptions of Caves

Gorman Cave

The entrance to Gorman Cave is in a bluff on the southern side of the Colorado River, 1520 m southeast of Gorman Falls in the east-central part of the Bend area (Figure 51). Gorman Cave is 920 m long and has an end-to-end extent of 690 m (Figure 52). The cave is nearly horizontal with approximately 10 m of internal relief. The main entrance is 5 m wide and 2.5 m high (Figure 53) and a small, second entrance, excavated in recent years along the east wall of the cave about 30 m from the original entrance, opens into a small depression at the base of a 10 m-high bluff.

Gorman Cave generally consists of a single south-trending passage averaging 5 to 9 m in width and 3 to 5 m in height (Figure 54). In places small crawlways, such as the 23 m-long Mouse Room, branch from the main passage, but most terminate shortly or form cutarounds. Constrictions in the main passage occur just inside the entrance, just north of the Big Room, in zones of collapse between Separation Lake and CO_2 Alley, and at the First Siphon. The main passage enlarges to 10 m wide by 8 m high at the Big Room, 210 m from the entrance. A lower-level complex containing an active stream provides an alternate route from the Bat Room to the Swiss Cheese and to holes in the floor in the collapsed zone at the Funnel.

Passages of the cave consist of a series of short, linear segments, the longest of which is a section of CO_2 Alley 90 m in length. The cave has numerous meanders, but it trends generally north and nearly perpendicular to the Colorado River.

The entire cave has formed within the calcitic facies of the Gorman Formation of the Ellenburger Group (Cloud and Barnes, 1946a). Large crystals of calcite line parts of the walls throughout the cave. Bedded chert is prevalent along the cave, but apparently it has had little effect on the development of passages.

A perennial stream in the southern section of the cave enters through the two siphons and then immediately flows into a small opening in the west wall of CO_2 Alley. The lower-level stream 180 m to the north is apparently the continuation of this water-course, but it is not possible to follow the stream north from the Funnel area where it passes through a constricted siphon. Apparently the stream discharges through a small spring just beneath the surface of the Colorado River where the gully from the entrance to the cave meets the river. Discharges of the stream in the cave and the spring are similar in magnitude.

There are several shallow pools along the length of the cave, some of which rarely dry up owing to continual replenishment by seepage. The amount of infiltrated water seeping into the cave is small and highly localized, and deposition of speleothems is presently minimal. However, many older deposits are found throughout the cave and were formed at a time when rates of infiltration were greater. Some deposits are massive, particularly those just north of the Big Room.

Much of the first half of the cave is floored with deposits of pebbles and gravel (Figure 54). These become interbedded with course sand, silt, and laminated clay in the central part of the cave (Figures 55 and 56).

Deposits of breakdown are few, especially in the first half of the cave where they only occur near the entrances. But, extensive breakdown occurs in areas of collapse extending from the Funnel to the Crack.

The entrance to Gorman Cave is approximately 50 m from the Colorado River and 6 m above normal pool-stage. The surface over the cave has been deeply dissected by the incision of the Colorado River (Figure 57). All drainage within 1-2 km of the river flows directly into the Colorado through short tributaries. There are at least two sets of fluvial terraces along both banks of the river in the vicinity of the cave entrance (Figures 57 and 58). These lie approximately 2.5 and 6 m above pool-stage of the river. Further descriptive, biological, historical, and bibliographical information on the cave may be found in Reddell and Estes (1962), Reddell (1964b, 1973) and Fieseler and others (1978).

Lemons Ranch Cave

Lemons Ranch Cave is part way up the side of a draw, 1675 m south-southeast of Gorman Falls and 850 m southwest of the entrance to Gorman Cave (Figure 51). The cave is 168 m long, 120 m in maximum extent, and 30 m in internal relief (Figure 59). The single entrance is a pit dropping 5.8 m into a small 2 m by 3 m room. To the east of the entrance are two chambers connected by steeply sloping, low passages. The easternmost and largest room is irregularly shaped and approximately 10-12 m in length and width. The mud-floored slope of the room leads to an 18 m-deep pit to the east from which no continuation has been found. The main passage of the cave trends north from the large room, initially as a crawlway and then as a 1 to 2 m-high, 2 to 3 m-wide corridor. About 50 m from the room this passage enlarges to 2 to 3 m in height and 3 to 5 m in width. These dimensions are maintained to the explored end of the cave except at one constriction in breakdown 80 m from the room. The passage terminates in a 5.5 m-high dome and blockage by flowstone. A crawlway above the flowstone, issuing strong currents of air, has

Figure 54: Main passage in Gorman Cave. Note thin- to medium-bedded units of limestone. Walls and ceiling exhibit scalloping and solutional pockets characteristic of modification by floodwater. Gravel on floor has been brought into cave by backflooding of the Colorado River.





Figure 55: Bank of sediment in Gorman Cave along east wall of main passage, approximately 6 m north of the Squeeze. Note laminated silt and clay in center of photograph and coarse sand and gravel in foreground. Hammer is 33 cm long. Turbulence of floodwater is evident from scalloping of deposit of silt and clay.



Figure 56: Bank of sediment in Gorman Cave along east wall of main passage, approximately 4 m north of Separation Lake. Note interbedded gravel and clay indicating variable conditions of energy during flooding of cave. Trowel is approximately 23 cm long.

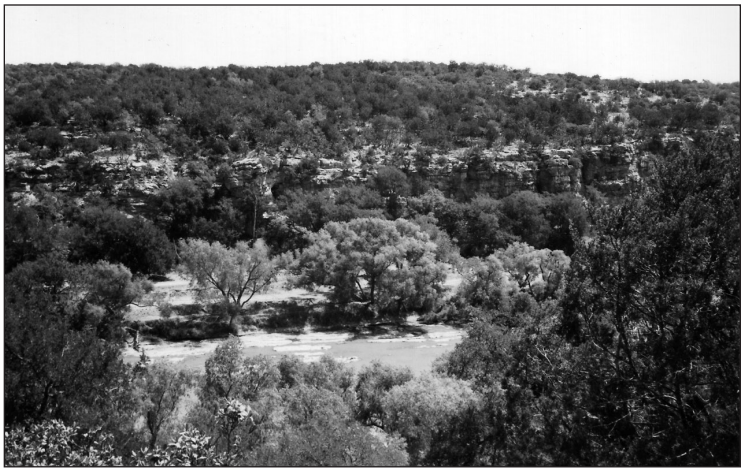


Figure 57: View northeast from top of bluff directly over Gorman Cave showing Colorado River and terraces, cliff in the Gorman Formation, and entrance to Canyon Wren Cave (dark opening approximately two-thirds of the way up from terrace to top of cliff, just left of center of photograph).



Figure 58: Terraces and bluff of limestone across Colorado River from Gorman Cave. Note two levels of terraces approximately 2.5 m and 6 m above the level of the river.

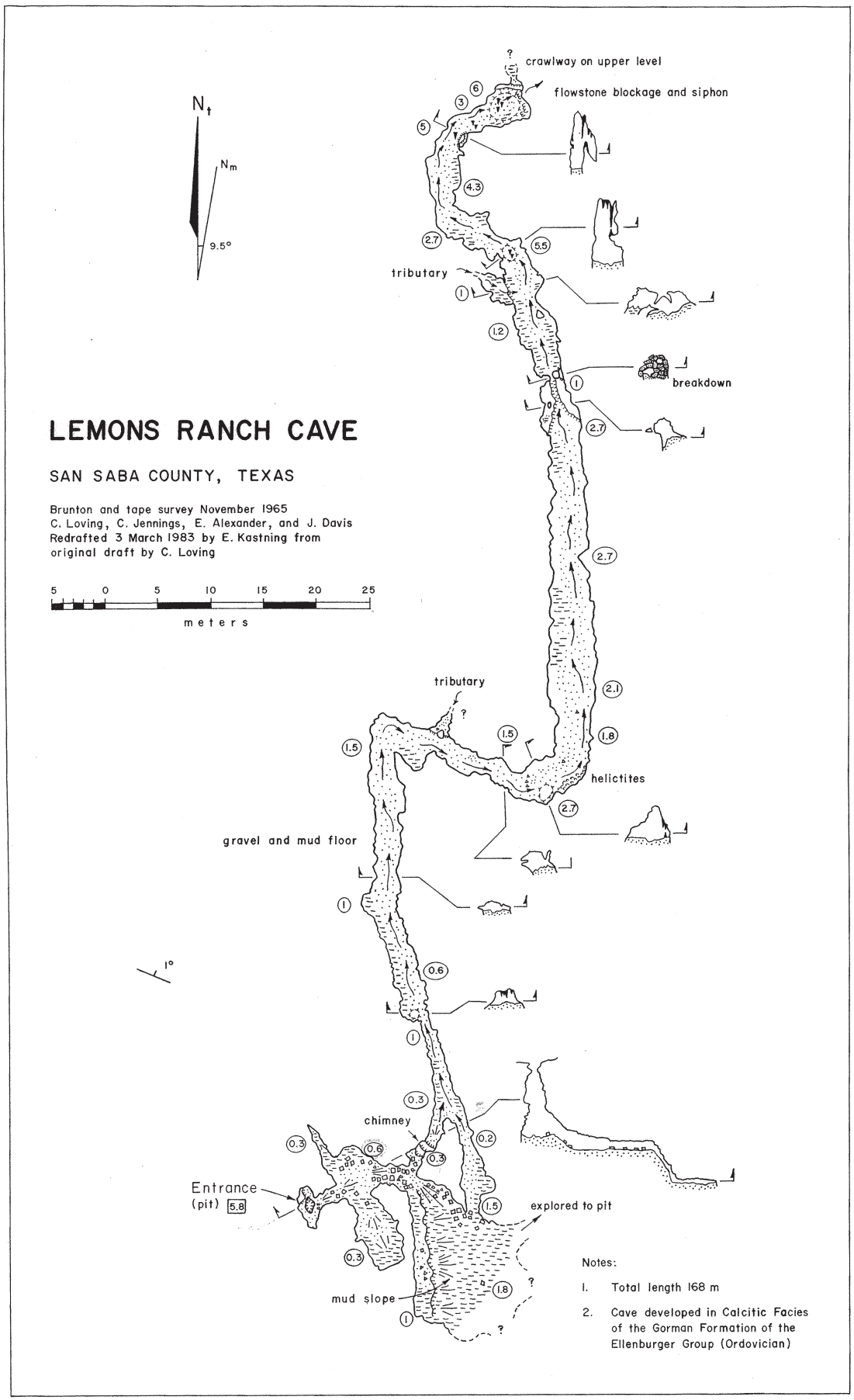


Figure 59: Map of Lemons Ranch Cave. Modified from Reddell (1973, p. 80).

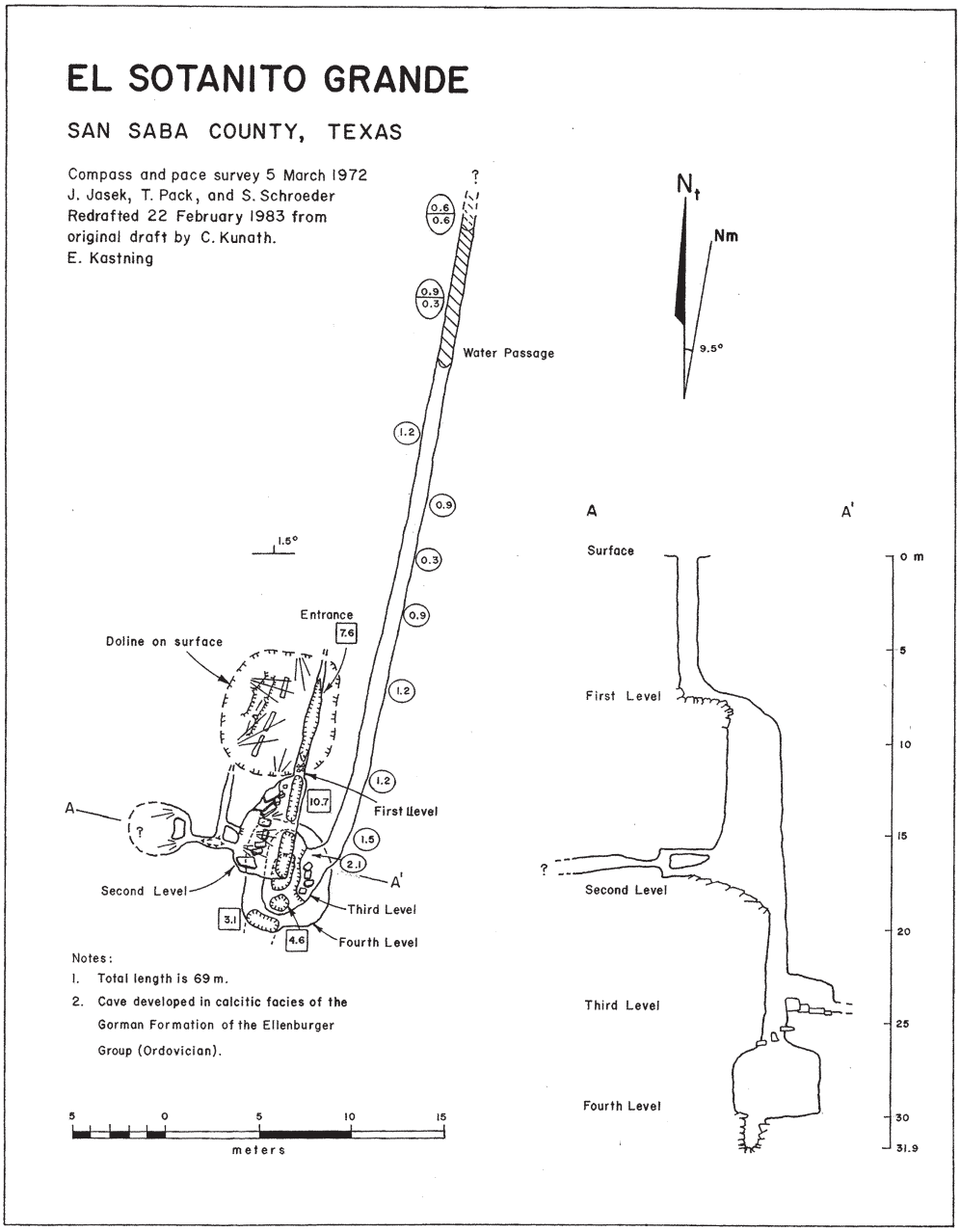


Figure 60: Map of El Sotanito Grande. Modified from Reddell (1973, p. 102).

been partially excavated and may lead to additional passages. Small alcoves and short side-passages occur locally along the cave, but these are no more than a few meters in extent.

Lemons Ranch Cave trends north from the entrance and consists of a series of linear segments up to 38 m long. Two abrupt, angular bends occur midway along the main passage where the northerly trend is offset by an easterly segment for 15 m.

Stratigraphically, the cave lies in the calcitic facies of the Gorman Formation. Most passages are conformable with the bedding, particularly along the main corridor, and generally follow the shallow dip to the north (Figure 59).

A narrow fluvial channel has been cut in the floor through most of the cave, but it remains dry except during periods of high precipitation and runoff when flow is to the north. Infiltration through overlying beds and seepage into the passages are moderate. Deposits of flowstone, stalagmites, stalactites, and helictites are locally profuse in the main corridor.

Terrigenous sediments generally include sand, silt, and clay. Clay is present in greater proportion here than it is in Gorman Cave and banks of clay line much of the stream-course. Collapse of the ceiling has produced considerable breakdown in the rooms at the south end of the cave. The single large room shows evidence of enlargement through stoping.

The lowest levels of the cave are approximately 8 m below the draw near the entrance to the cave and 45 m above the Colorado River. Surface drainage is to the northeast through moderately dissected terrain.

El Sotanito Grande, Yates Goat Cave, Barnes Bat Cave

El Sotanito Grande, Yates Goat Cave, and Barnes Bat Cave are respectively approximately 460 m, 580 m, and 1130 m south of Lemons Ranch Cave (Figure 51). The entrance of El Sotanito Grande is a pit in a shallow doline at the end of a small draw. The cave consists of a series of four successive drops of 7, 11, 6, and 5 m, and a total depth of 32 m (Figure 60). At the base of the second drop is a small room, 3 m wide and 5 m long, with short, low passages extending to the west. A 1 to 2 m-high passage leads north-northeast from the east wall at the base of the third drop and continues 35 m to where it is nearly filled with water. A room 5 m in diameter occurs at the bottommost level and a 3 m-deep pit at its south end is unexplored. Total horizontal length of the cave is 69 m. Following storms the cave must take in considerable runoff from the draw at the entrance.

The entrance to Yates Goat Cave lies in a shallow doline about 5 m long and 11 m wide. Although only 12 m in horizontal extent, the cave is 20 m deep and two levels branch to the east and west at 5 m and 18 m below the surface (Figure 61). Neither of these extends more than 6 m before being filled with sediment. The lowest

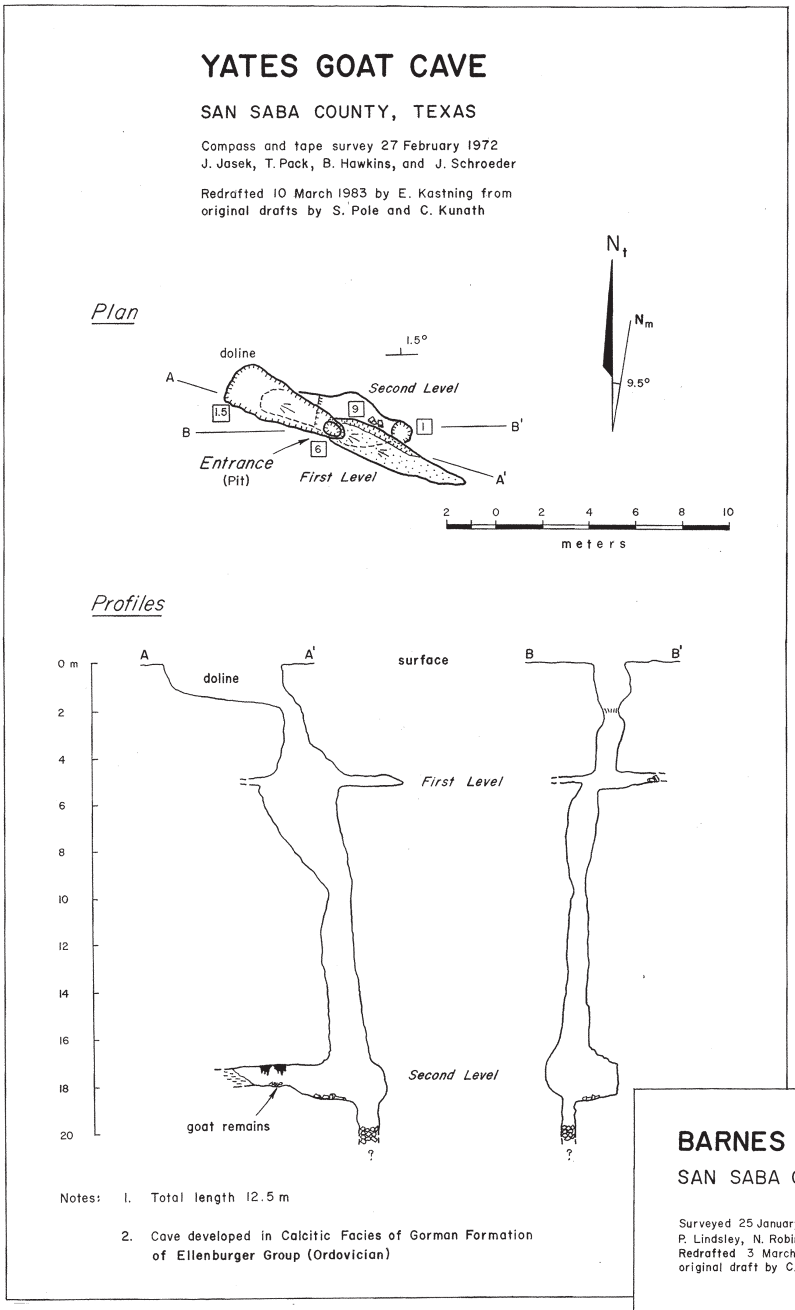


Figure 61: Map of Yates Goat Cave. Modified from Reddell (1973, p. 120).

point in the cave is plugged with fragments of rock. The cave is generally dry,but some speleothems occur in the lowest, horizontal passage.

The entrance to Barnes Bat Cave is on the northeastern slope of the valley of Spicewood Creek, and consists of a 10 m-deep pit that opens into a room 12 m wide, 15 m long, and 1.5 m high (Figure 62). Two small side-passages branch to the northeast and southwest, but the main passage of the cave extends from the southeast corner of the room and continues as a series of straight segments forming a 140 m-long loop to the south. Half way along the loop the passage enlarges to a width of 3 to 6 m and a height of 2 to 5 m (Big Bat Room) and continues at these dimensions for approximately 30 m, to where its cross-section is reduced to a low, filled area, perhaps lying beneath a doline. Total length of the cave is 158 m, and total depth is 10 m. Apparently the cave floods during high runoff in Spicewood Creek. An intermittent stream flows from the terminus of the main passage, through the Big Bat Room, and into a small passage branching to the southeast. Several small pools are present in the Big Bat Room.

All three caves have developed in the calcitic member of the Gorman Formation, and horizontal passages are conformable with beds dipping 1° to 2° to the north (Figures 51, 60, 61, and 62). Major levels in El Sotanito Grande and Yates Goat Cave correspond in elevation suggesting lithologic control (discussed below). Passages of all three caves are floored with mud and silt, and in the two deep caves, gravel and cobbles choke the lowest levels. Breakdown is limited to the horizontal shelves forming the intermediate levels in El Sotanito Grande and Yates Goat Cave and to the entrance-room of Barnes Bat Cave. The three caves have formed under the flanks of draws whose floors generally lie between 6 and 24 m below their entrances.

Clark's Branch Well Cave

The entrance to Clark's Branch Well Cave is alongside the bed of Spicewood Creek 2760 m south-southeast of Gorman Falls and 180 m west-southwest of Barnes Bat Cave (Figure 51). The cave is the second longest in the area with a length of 335 m and an end-to-end extent of 255 m (Figure 63). Its vertical extent is 12 m. The entrance is a 7.6 m-deep pit into a small room 5 m in diameter which connects to a second, similar room to the south. A sloping passage extends from this room to the main

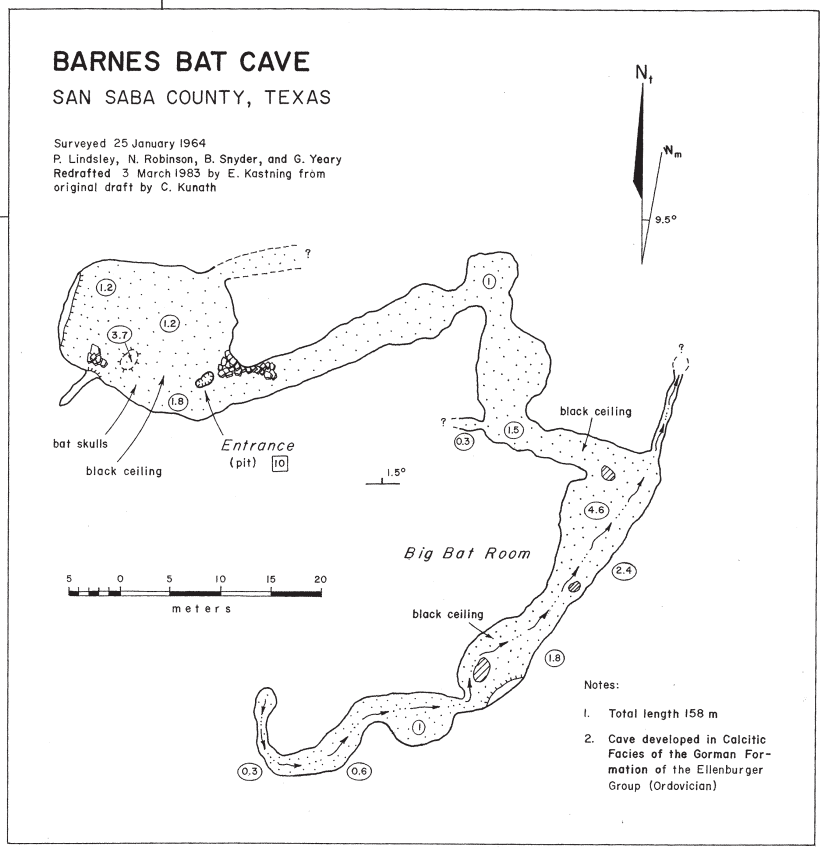


Figure 62: Map of Barnes Bat Cave. Modified from Reddell (1973, p. 16).

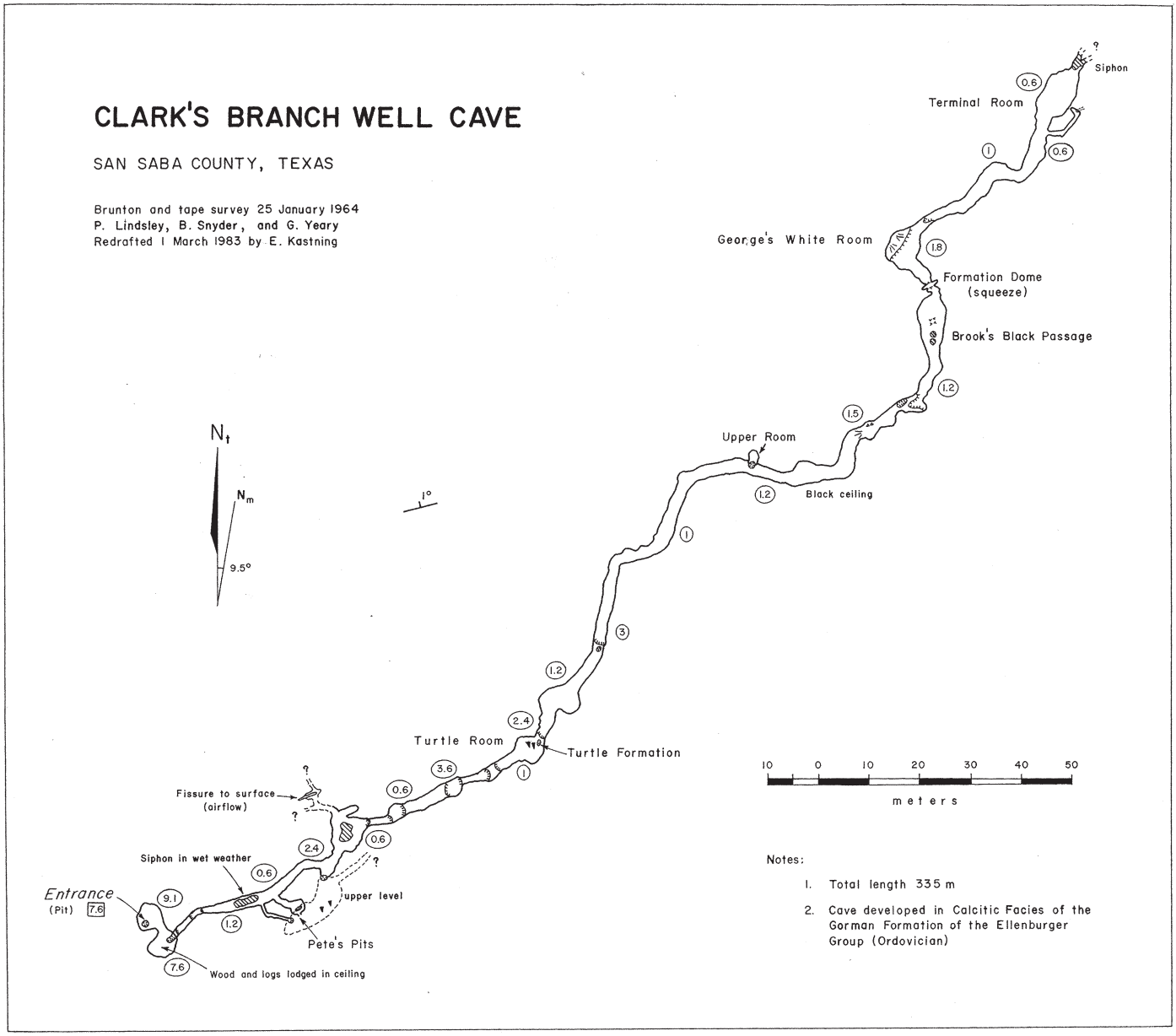


Figure 63: Map of Clark's Branch Well Cave. Modified from Reddell (1973, p. 31).

passage of the cave which trends horizontally to the northeast in a meandering fashion. Along its course this corridor is generally 1 to 2 m high and 3 to 5 m wide and consists of linear segments generally 15 m or less in length. Short side-passages in the southwestern end of the cave lead to an upper-level room to the southeast, and to a vertical fissure to the northwest that passes a considerable flow of air. There are two short side-passages in the northeastern part of the cave, but these end abruptly. In places the main passage widens up to 7 m and contains shallow pits, domes, and short drops.

Clark's Branch Well Cave lies within the calcitic facies of the Gorman Formation and is conformable with beds dipping 1° to the north-northwest (Figure 63). The cave is generally very wet, with shallow pools along its entire length. One such pool forms a siphon 20 m from the entrance and another nearly blocks the explored and mapped end of the cave. The cave apparently floods during high runoff in Spicewood Creek. Gravel and sand floor much of the cave, and there are relatively few deposits of breakdown. Speleothems occur in several places, but they are not exceptionally large. The cave is generally close to the local baselevel at Spicewood Creek in an area of moderate to high dissection of the surface.

Gorman Fissure System

The Gorman Fissure System consists of four fissure-like caves (Crevice Cave, McLarrin Flowstone Fissure, Toad Frog Falling Floor Fissure, and Coon Cave) located on a topographic bench 140 to 213 m southwest of the Gorman Cave entrance (Figures 51 and 64). The caves are generally developed as single fissures or crevices, 5 m in horizontal length and 5 to 17 m in depth (Figure 65). Each has a narrow slit-like entrance, typically 0.3 to 0.7 m wide, that drops vertically into a fissure below (Figure 66). The widths of the caves (0.3 to 0.5 m) remain uniform with depth but taper horizontally.

The caves lie within the calcitic facies of the Gorman Formation. Lithic fragments and surficial debris litter the bottoms of the fissures as well as ledges at various levels. This loosely consolidated material allows meteoric water to infiltrate to lower depths during storms, but the caves remain dry during much of the year. Some deposition of flowstone and helictites are found in McLarrin Flowstone Fissure and Toad Frog Falling Floor Fissure. The caves apparently drain northward to the Colorado River which lies about 40 m below the elevation of their entrances (Figure 64).

Bluff-Caves

Five small caves occur in the cliffs along the Colorado River, including three in Lampasas County (Canyon Wren Cave, Soot Cave, Hole in the Wall) and two in San Saba County (Skull Cave, Climbup Cavelet). Canyon Wren Cave, approximately 200 m downstream and across the river from Gorman Cave (Figures 51, 57, and 64), lies approximately 15 m above the river and consists of two linear, horizontal segments of passages totalling 20 m in length (Figure 67).

Soot Cave is located directly across the river from the large terrace of travertine at Gorman Falls and 12 m above the river (Figures 68 and 69). It consists

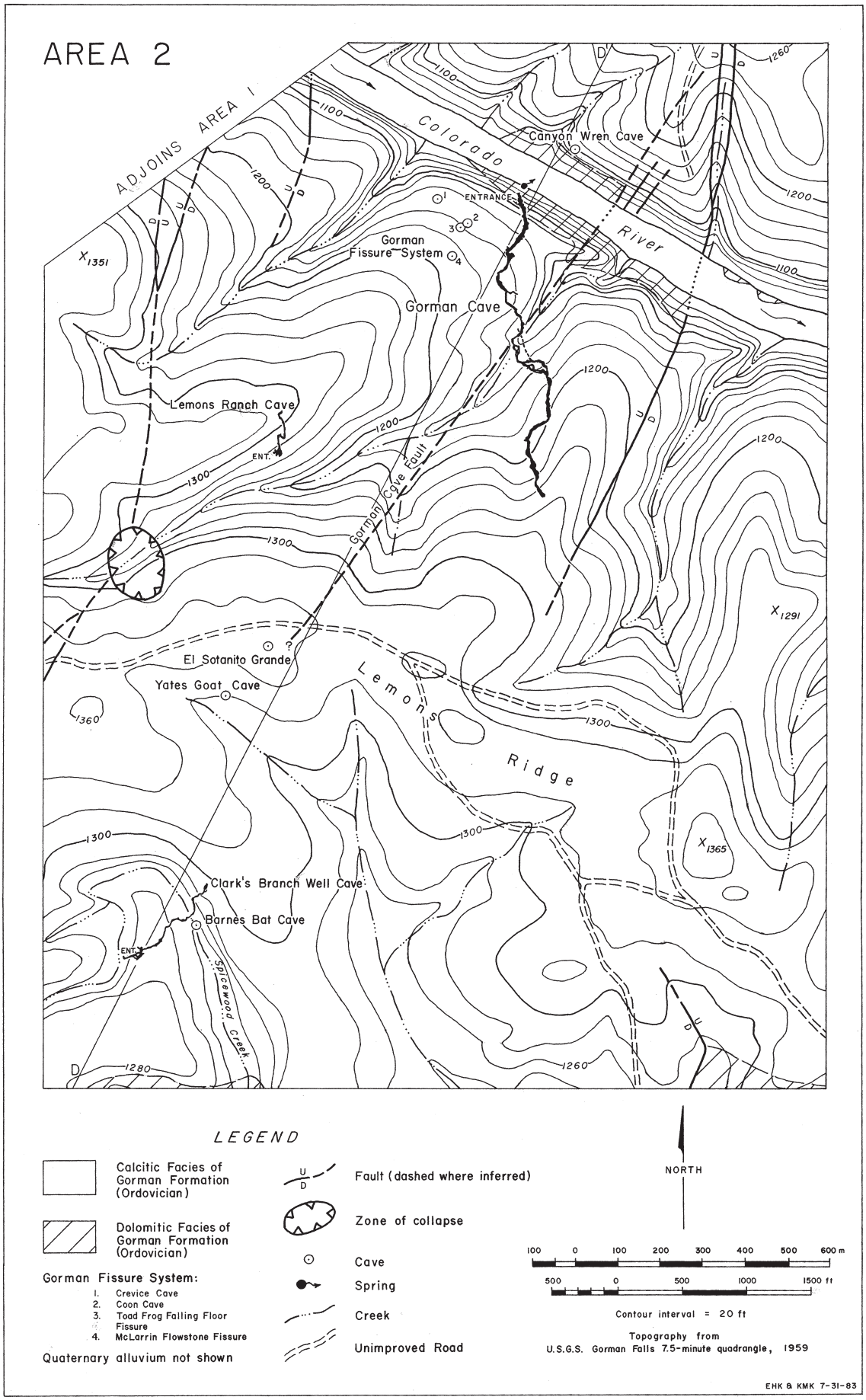


Figure 64: Detailed map of Area 2 of Figure 51 showing topography, caves, and structure. Geologic section D-D' is shown in Figure 91. Geology modified from Cloud and Barnes (1946a, plate 9).

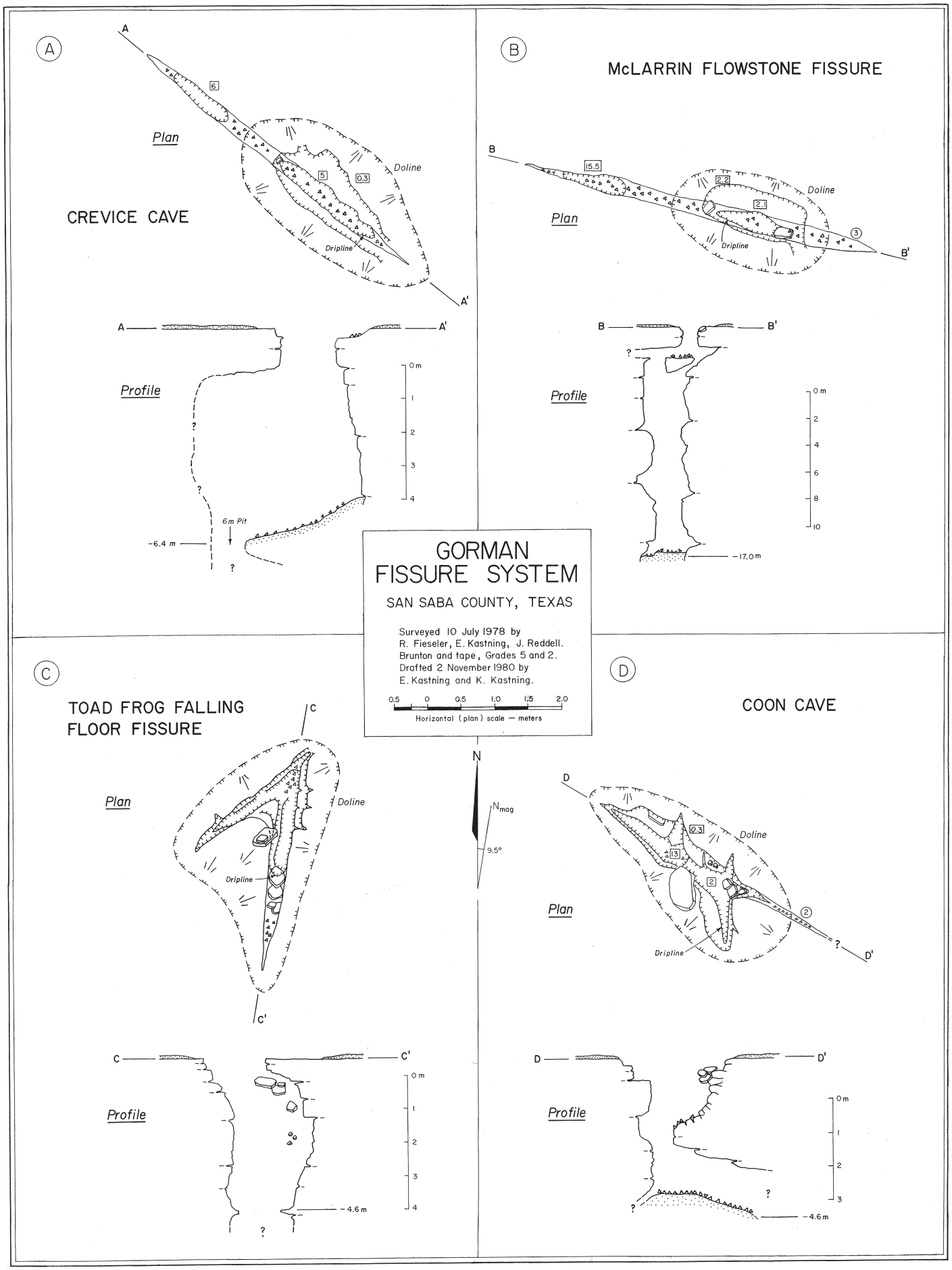


Figure 65: Maps of caves of the Gorman Fissure System.



Figure 66: Entrance of Toad Frog Falling Floor Fissure, Gorman Fissure System, showing thin-bedded limestone and control by fractures. View is to the south.

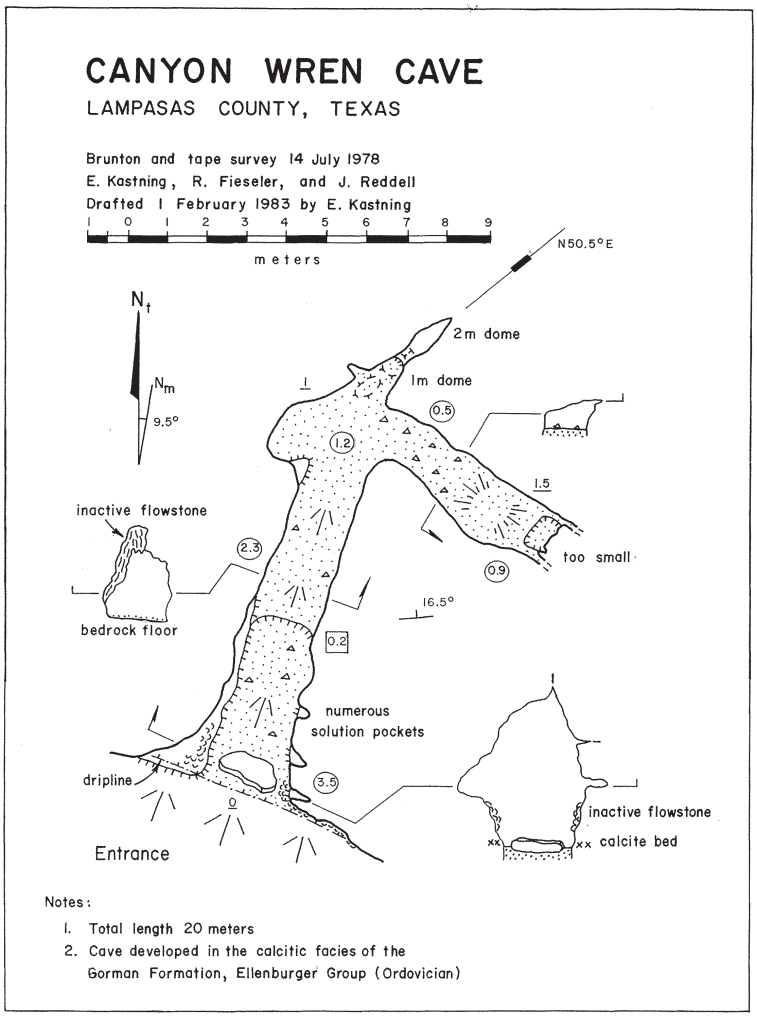


Figure 67: Map of Canyon Wren Cave.

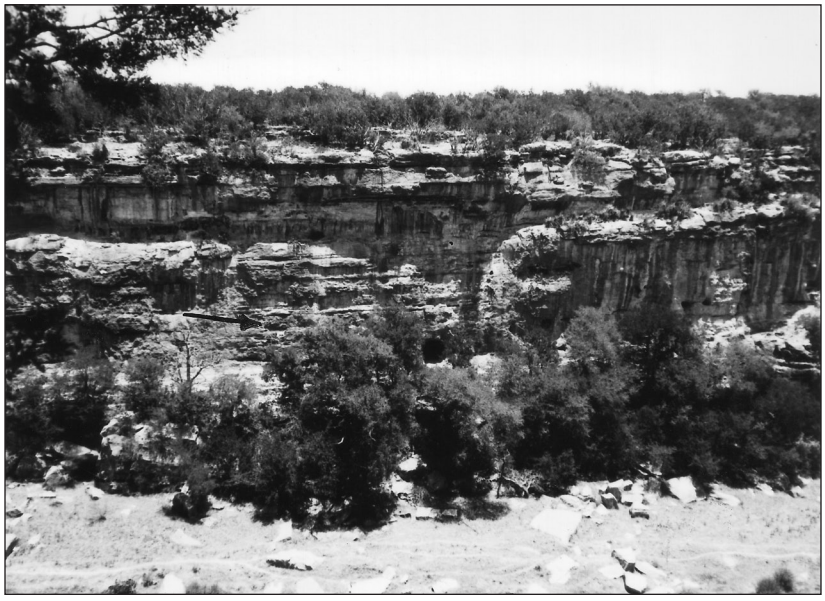


Figure 68: View northeast from top of deposit of travertine near Gorman Falls, showing bluff of Gorman Falls, showing bluff of Gorman Formation with entrances of Soot Cave (center) and Hole in the Wall (center left at arrow).

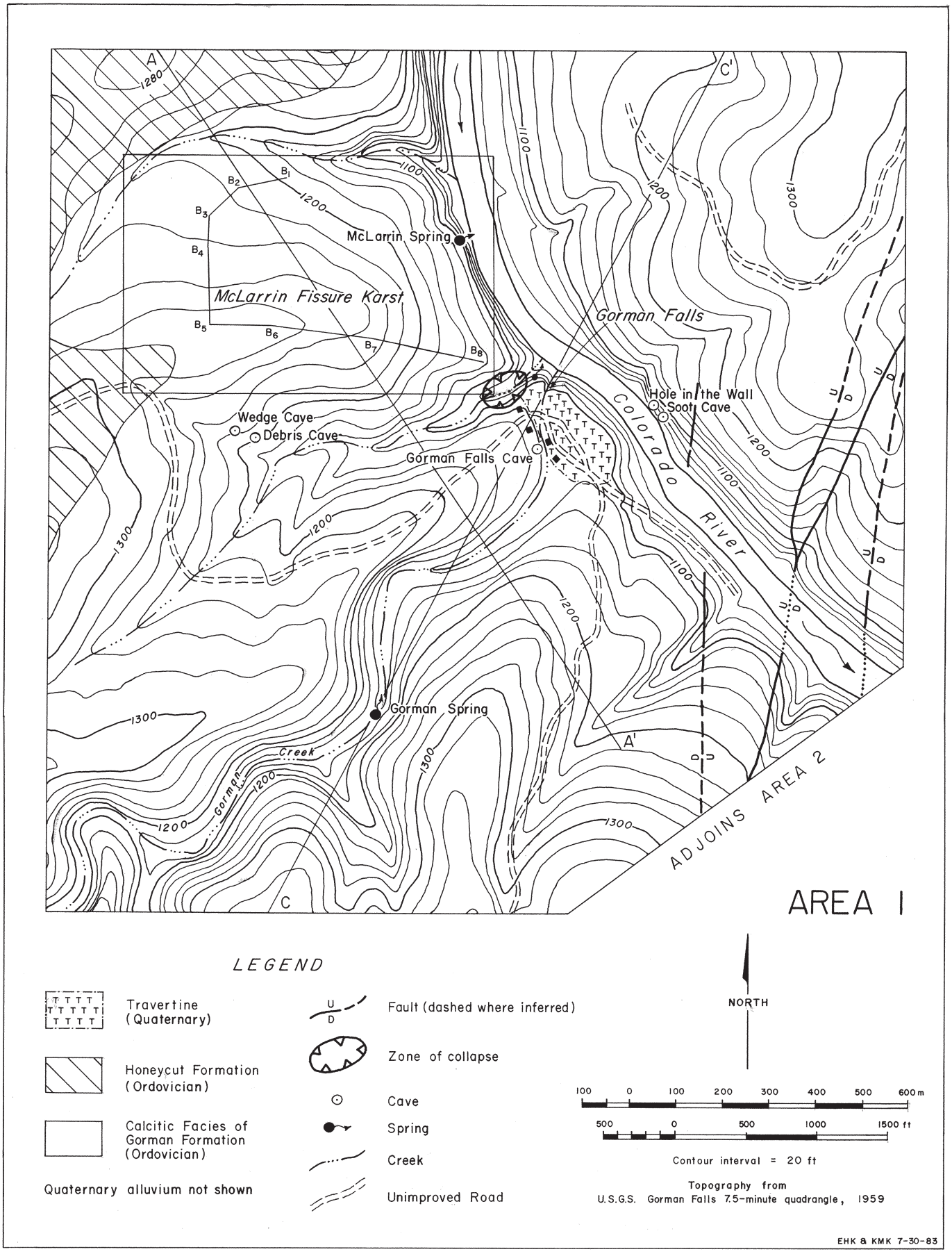


Figure 69: Detailed map of Area 1 of Figure 51 showing topography, caves, structure, and deposit of travertine. Area of McLarrin Fissure Karst is shown in Figure 73. Geologic sections A-A', B-B', and C-C' are shown in Figures 89, 74, and 90 respectively. Geology modified from Cloud and Barnes (1946a, plate 9).

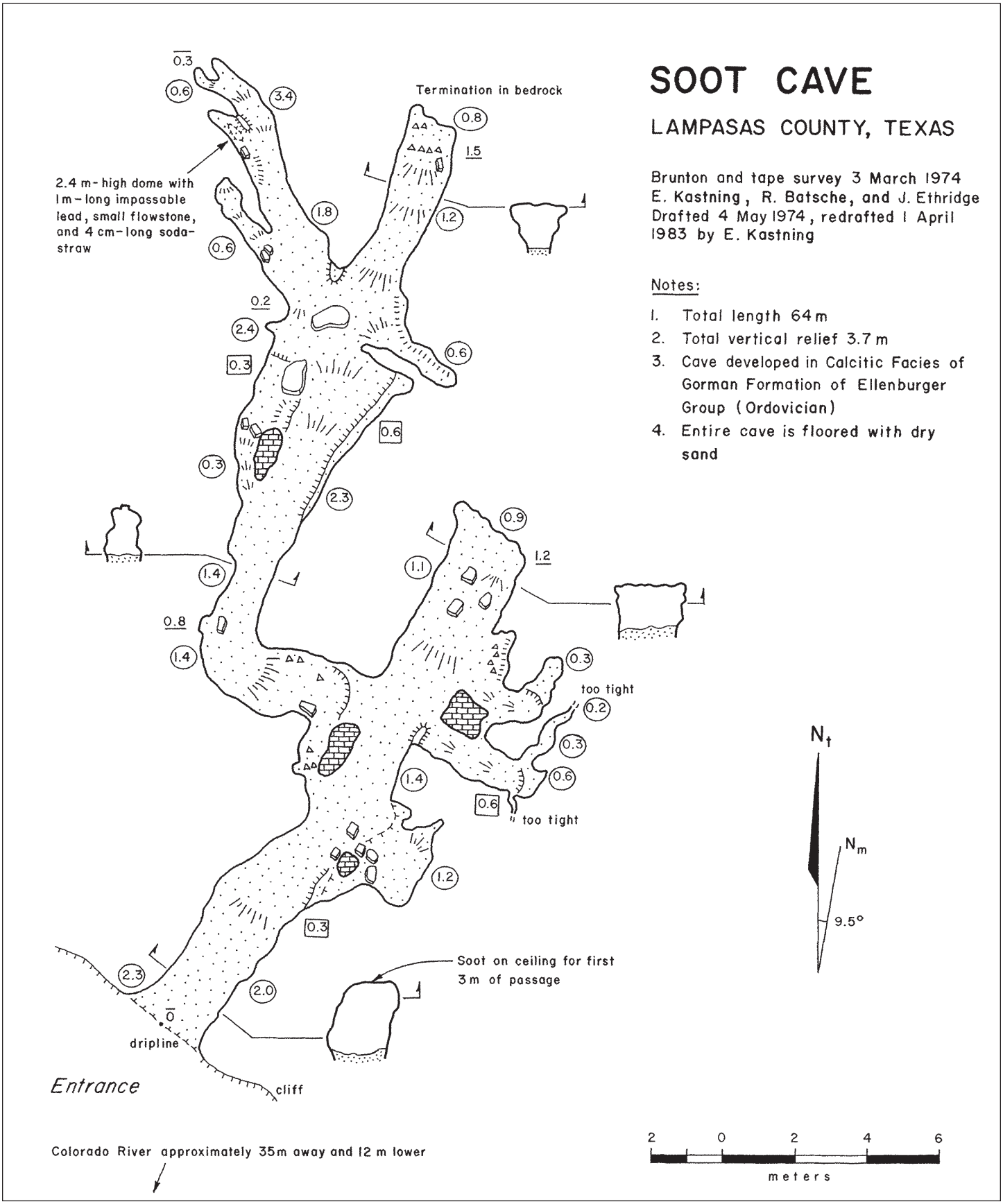


Figure 70: Map of Scot Cave. Modified from Kastning (1974, p. 128).

of two northeast-trending, horizontal passages intersected by several shorter, northwest-trending passages and has a total length of 64 m (Figure 70). Hole in the Wall, approximately 30 m to the northwest of Soot Cave (Figures 68 and 69), is a crawlway, opening 5 m up the cliff face and extending 11 m to a 2.3 m-high dome (Figure 71).

Skull Cave and Climbup Cavelet are in the cliff to the west of the Colorado River approximately 340 m northwest of Gorman Falls. These small caves, approximately 15 m above the river, are crawlways 14 m and 5 m in length respectively, with ceiling heights of 0.2 to 1.5 m (Figure 72; see Figure 73, map of McLarrin Fissure Karst, for location of caves).

All five bluff-caves are developed within the calcitic facies of the Gorman Formation. They are dry, relict features, are floored with sand, and contain small, inactive deposits of speleothems. Most are formed along vertical joints, but Skull Cave is guided by a joint plane dipping 39.5° to the southeast (Figure 72).

Northwest Caves

Copperhead Cave, 180 m west-northwest of Gorman Falls on the east flank of the uplands containing the McLarrin Fissure Karst (Figures 73 and 74), consists of a single, horizontal passage, 18 m in length, extending north from a 3 m-deep doline (Figure 75). The passage varies in width from 0.7 to 6 m and in height from 1 to 3 m. Surficial debris and lithic fragments litter the slope to the entrance; however, the terminal chamber is floored by a deposit of silt and clay.

El Sotano Lodoso, 61 m west of Copperhead Cave (Figures 73 and 74), is 33 m in length and 28 m deep and is one of the deepest caves in the Bend area (Figure 76). A 10 m-deep pit drops to a debris-covered slope along an east-trending passage which drops an additional 7 m to the top of a broad room, approximately 8 m in diameter and 2 m high, that slopes downward to the southeast. After an additional 9 m of depth, the room terminates in a small pit dropping 2 m into the lowest chamber, a northeast-trending room, 6 m long, 2 m wide, and 2 m high. The cave is floored by lithic fragments, some blocks of breakdown, and a silty-clayey floor. Some small deposits of flowstone are present.

Gorman Sink, 150 m west of El Sotano Lodoso (Figures 73 and 74), is a single chamber extending south from the bottom of a 2.5 m-deep collapsed doline (Figure 77). This room, 15 m long, 8 m wide, and 2.5 m high, is floored with breakdown, lithic fragments, and some silt and clay.

Debris and Wedge Caves lie alongside a small draw, 100 m south of the McLarrin Fissure Karst (Figure 51). Debris Cave consists of a northeast-trending crawlway with a short crawlway branching to the west near its end (Figure 78). The cave is 14 m long and averages 1 to 2 m in width and 0.5 to 1 m in height. It is floored with lithic fragments, silt and clay, and organic debris. Wedge Cave extends 16 m northeast from an elongated, collapsed doline (Figure 79). It begins as a 1 m-wide, 3 m-high crawlway and widens to a low, flat room at its inner end. Breakdown, lithic fragments, and silt and clay litter the cave, and the passage terminates in a plug of flowstone.

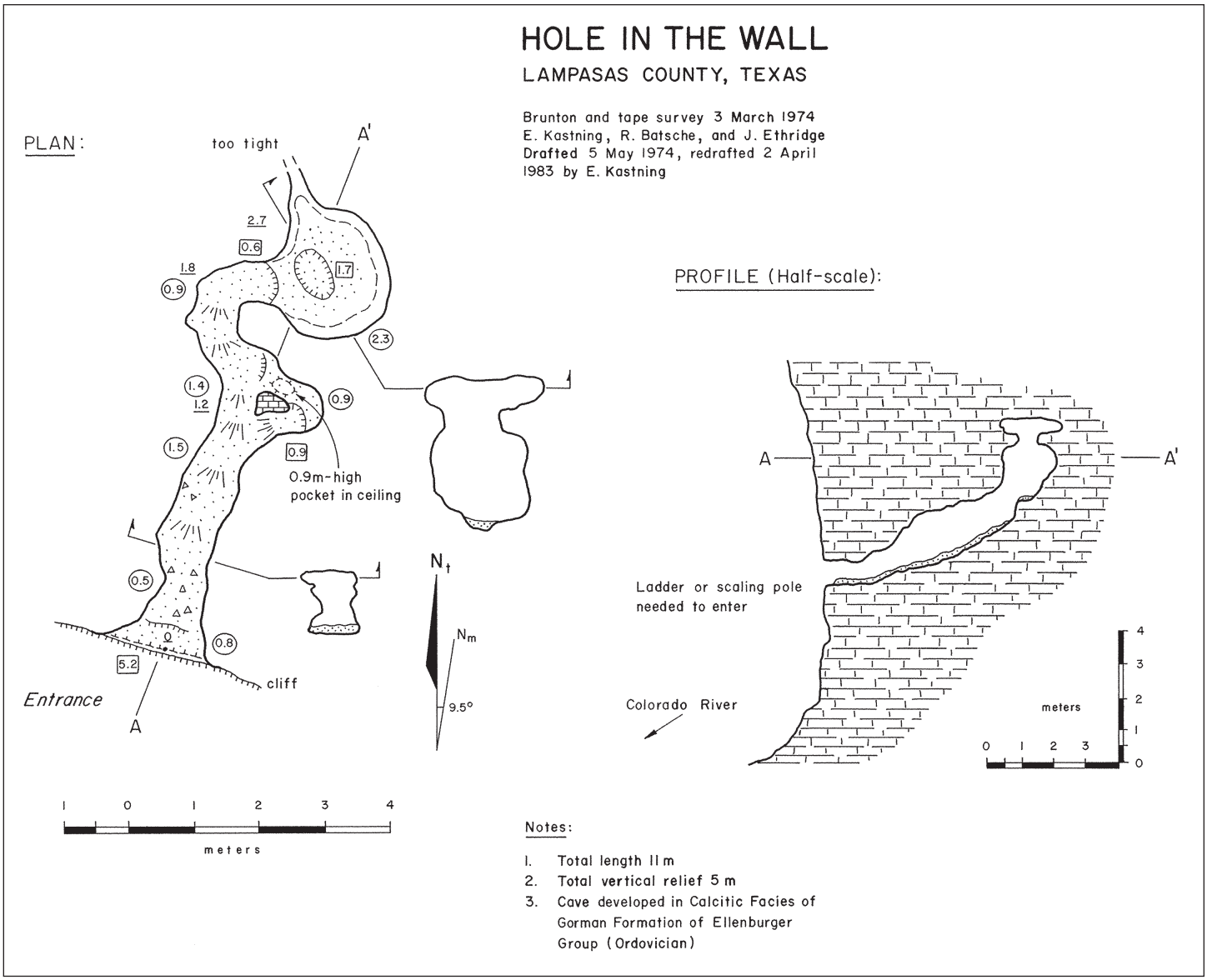


Figure 71: Map of Hole in the Wall. Modified from Kastning (1974, p, 129).

The five northwest caves described above are within the calcitic facies of the Gorman Formation. The sixth cave, Blue Haw Cave, in the northwest corner of the Bend area and on the northwest valley-wall of Tie Slide Creek (Figure 51), is developed in the Honeycut Formation. The cave consists of a 12 m-long passage that averages 0.6 m wide and 1.5 m high except at its entrance where the cave forms a large shelter, 8 m wide and 4 m high (Figure 80). The cave is developed along a joint that dips 75° to the northeast. It is floored with lithic fragments and silt and clay and contains some inactive deposits of flowstone.

Gorman Falls Cave

Gorman Falls Cave is 30 m west of Gorman Creek and approximately 140 m southwest of Gorman Falls (Figure 69). The entrance, an 8.5 m-deep pit, drops into a south-trending passage which descends an additional 6 m into a 19 m-high, domed room, followed by a third drop of 6 m into the main passage (Figure 81). This low-level passage, 1 to 5 m wide and 3 to 6 m high, extends south and northwest from this point as a series of linear segments and terminates in siphons at each end. Total length of the cave is 82 m and vertical relief is about 27 m. It is developed in the calcitic facies of the Gorman Formation, and contains a swiftly flowing stream entering the south end of the cave through an artesian spring, flowing along the entire length of the lowest passage, and passing through the northern siphon. Presumably this water emerges at a small spring visible at the base of the deposits of travertine 60 m northwest of Gorman Falls (Figure 69).

Caves in Travertine

Two small caves are developed within the large deposit of travertine at Gorman Falls. The largest of these, Gormlette Waterfall Cave (Figure 82) is several meters southeast of the falls and

about 15 m above the level of the Colorado River. Two entrances at the east end of the cave lead into a room 6 m in diameter and 1.5 m in height. A passage 0.6 to 2 m wide and 3 to 5 m high leads 6 m to the west to another room 5 m long, up to 5 m wide, and 1.3 m high. At the west end of this room is a third entrance opening onto the cliff near Gorman Falls. A 6 m-long lower passage leads north from the last room, but terminates before reaching the cliff. Total length of the cave is 23 m, and it is dusty and dry, but well decorated with dams of travertine, flowstone, and stalactites.

Mini Cave, another cave within the same travertine deposit, lies directly below the Gorman waterfall, approximately 12 m above the Colorado River. It is entered through a 0.5 m-diameter crawlway and consists of a room 8 m long and up to 2 m high with a 0.3 m-deep pool of water covering the floor (no map available). Small crystals-of calcite adorn the ceiling of the cave.

Caves of the McLarrin Fissure Karst

The McLarrin Fissure Karst includes 18 fissure-like caves (Walkby and Blacktail Pits; X, Meander Tube, Offset, Honeycombless, Two Part, Cobweb, Fern, King,

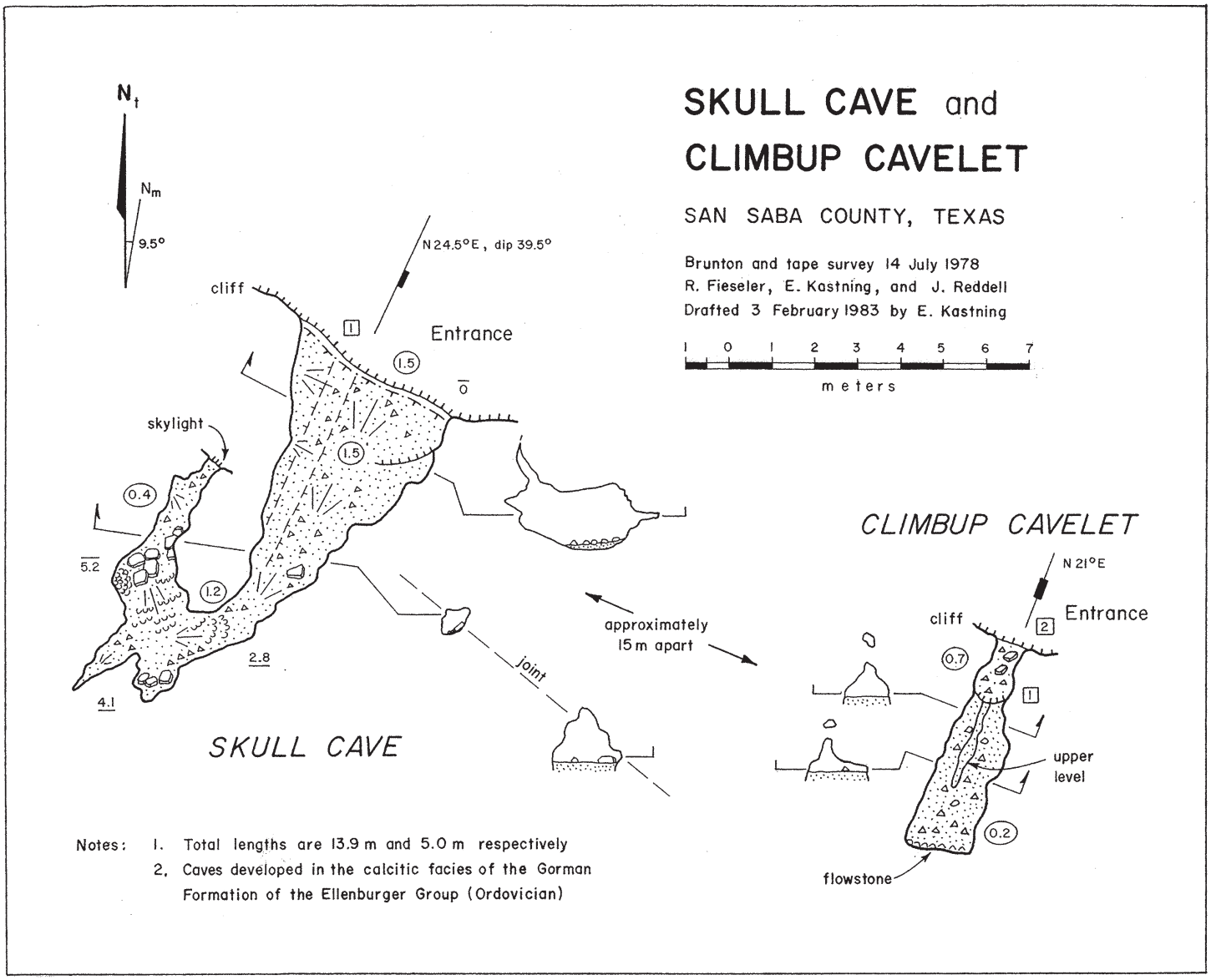


Figure 72: Top of Small Cave and Climbup Cavelet.

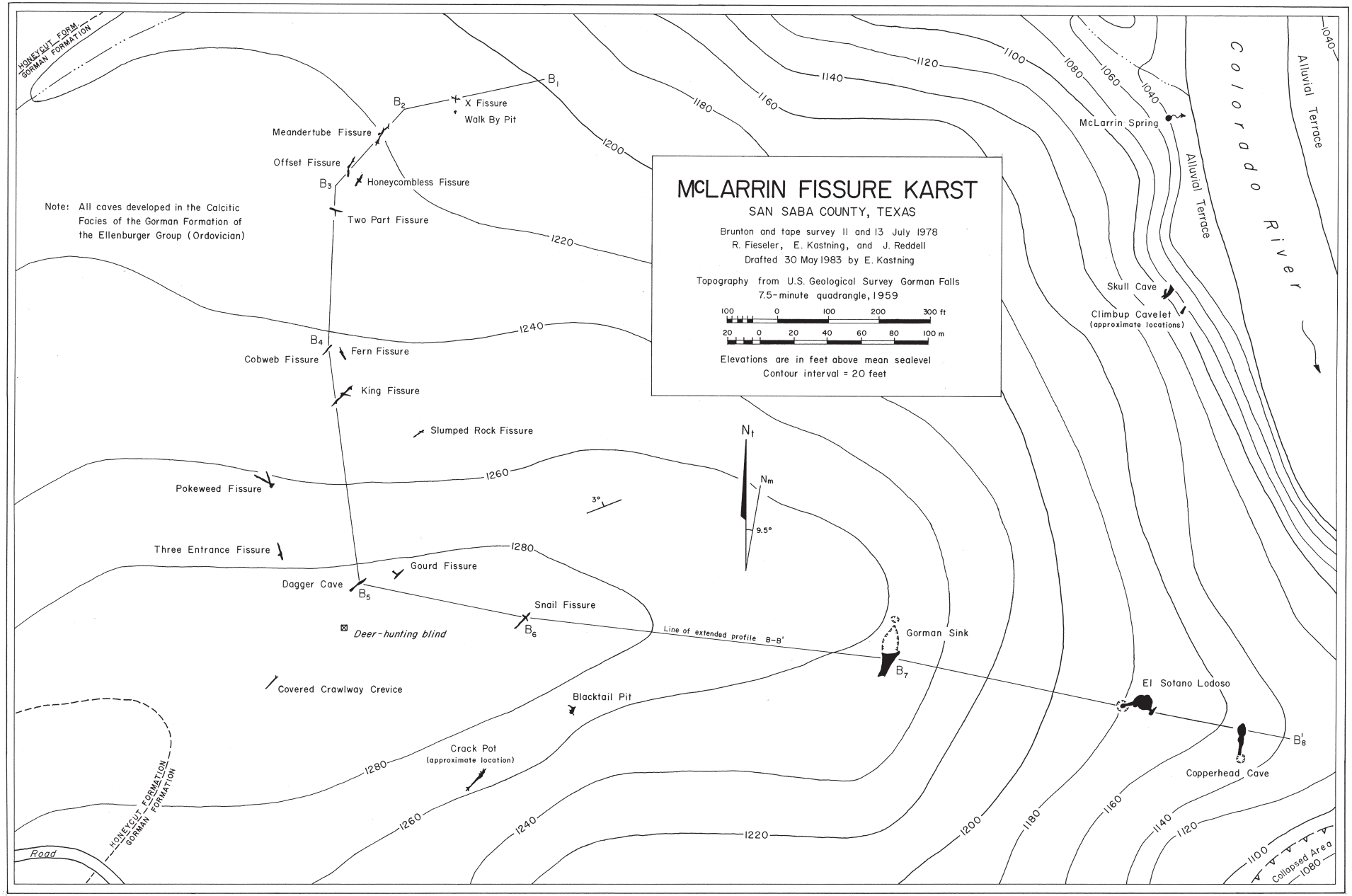


Figure 73: Topographic map of the McLarrin Fissure Karst, showing caves and geology. For location see Figure 69. Section B-B' is shown in Figure 74.

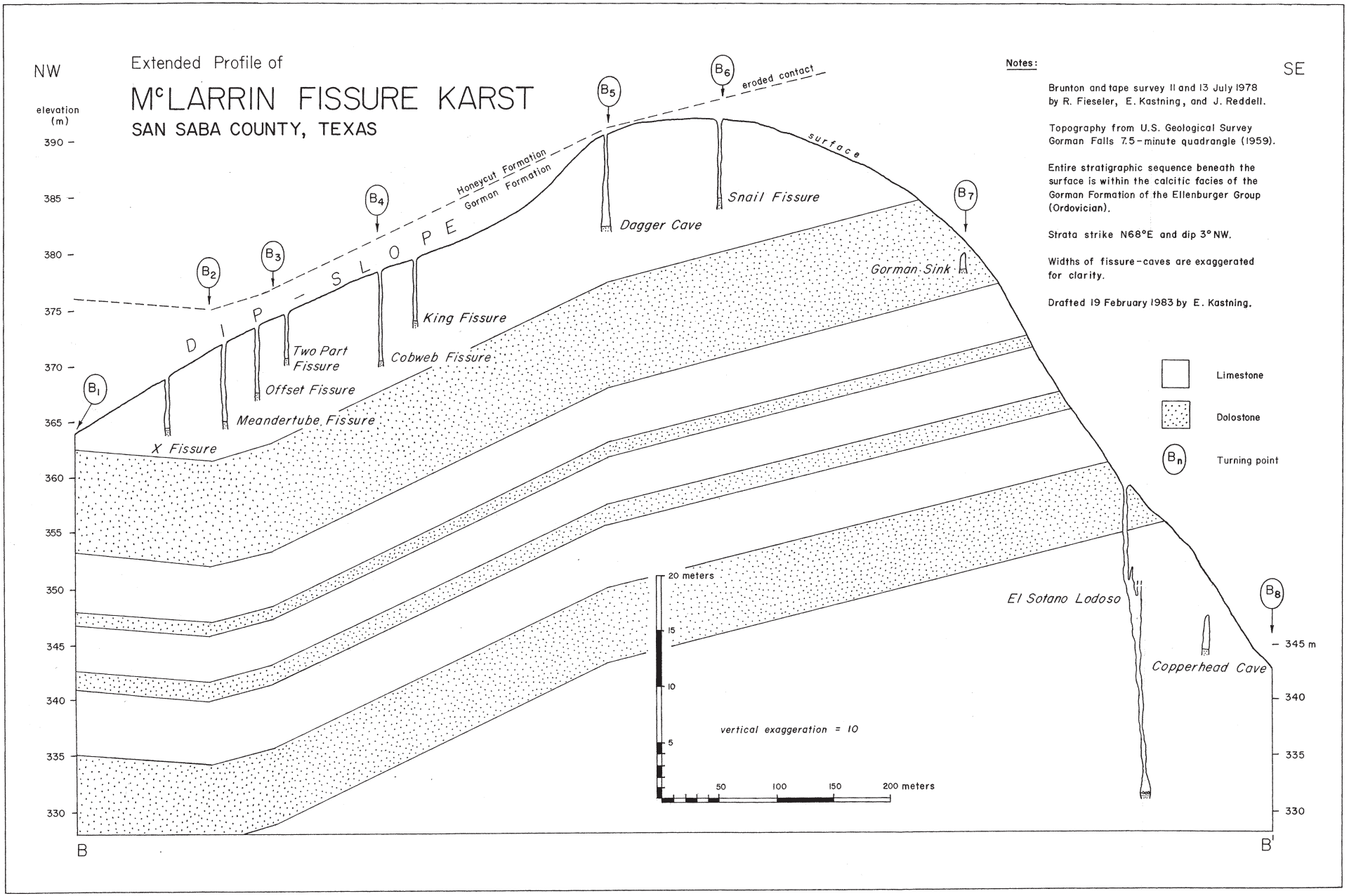


Figure 74: Geologic section B-B' of the McLarrin Fissure Karst, showing caves and stratigraphy. For location, see Figures 51, 69, and 73. See also Figure 89. Cross-sections of Gorman Sink, El Sotano Lodoso, and Copperhead Cave are based on maps by Schroeder, Jasek, Hawkins, and Pack in Reddell (1973).

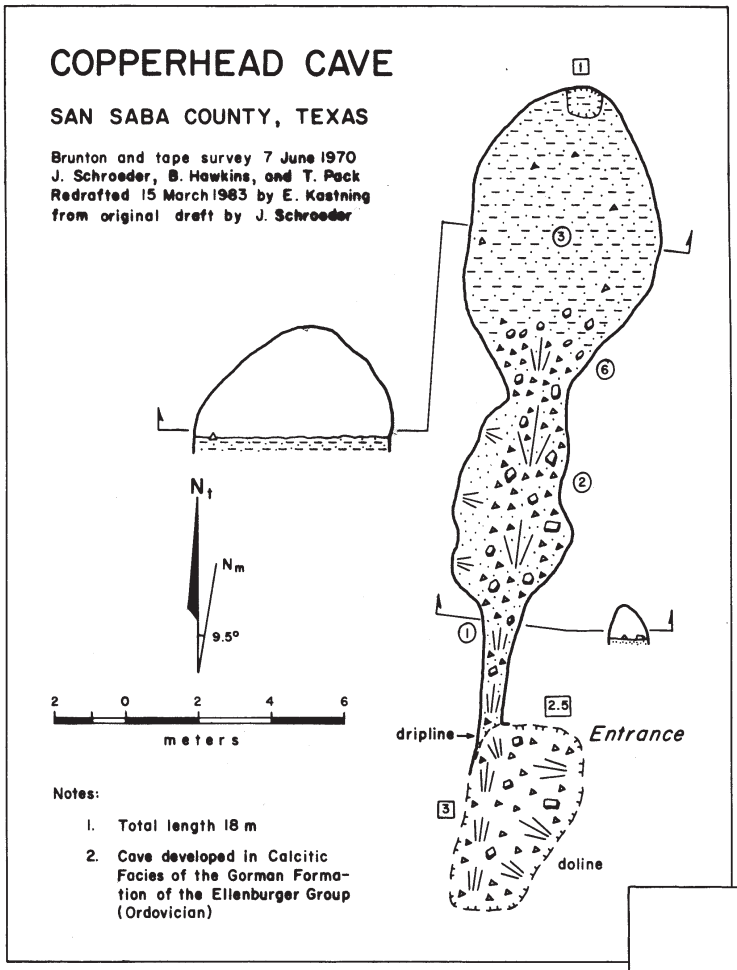


Figure 75: Map of Copperhead Cave. Modified from Reddell (1973, p. 34).

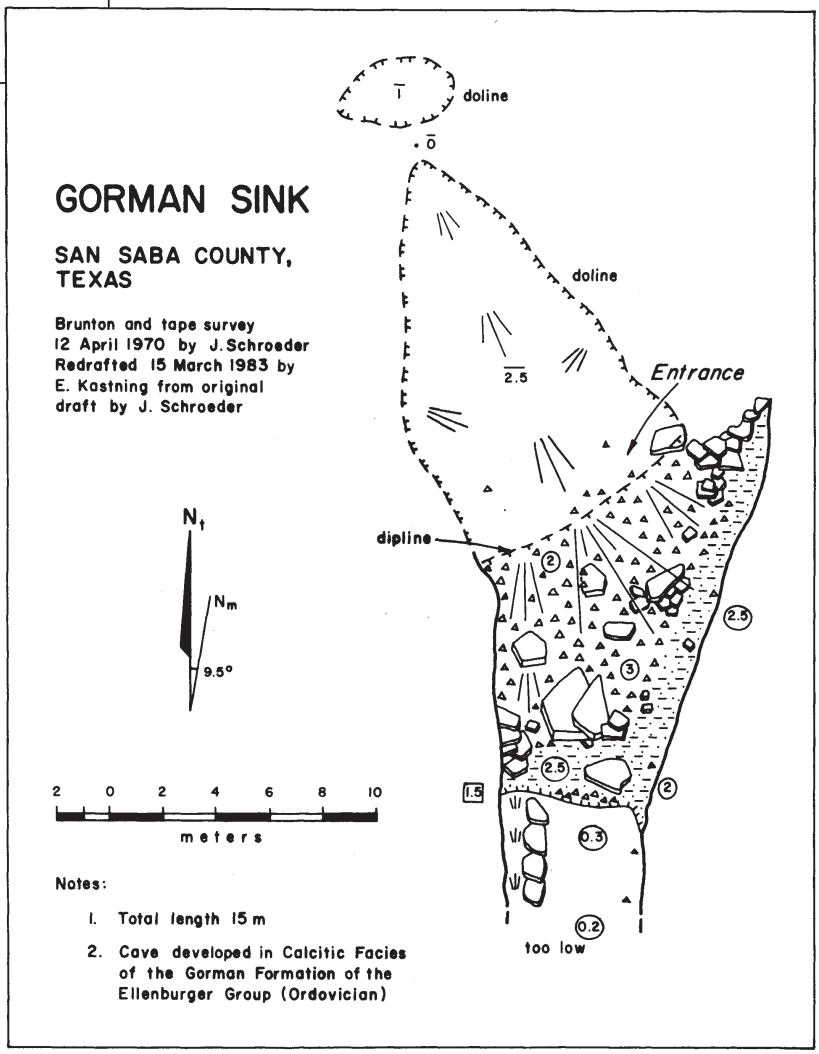


Figure 77: Map of Gorman Sink. Modified from Reddell (1973, p. 60).

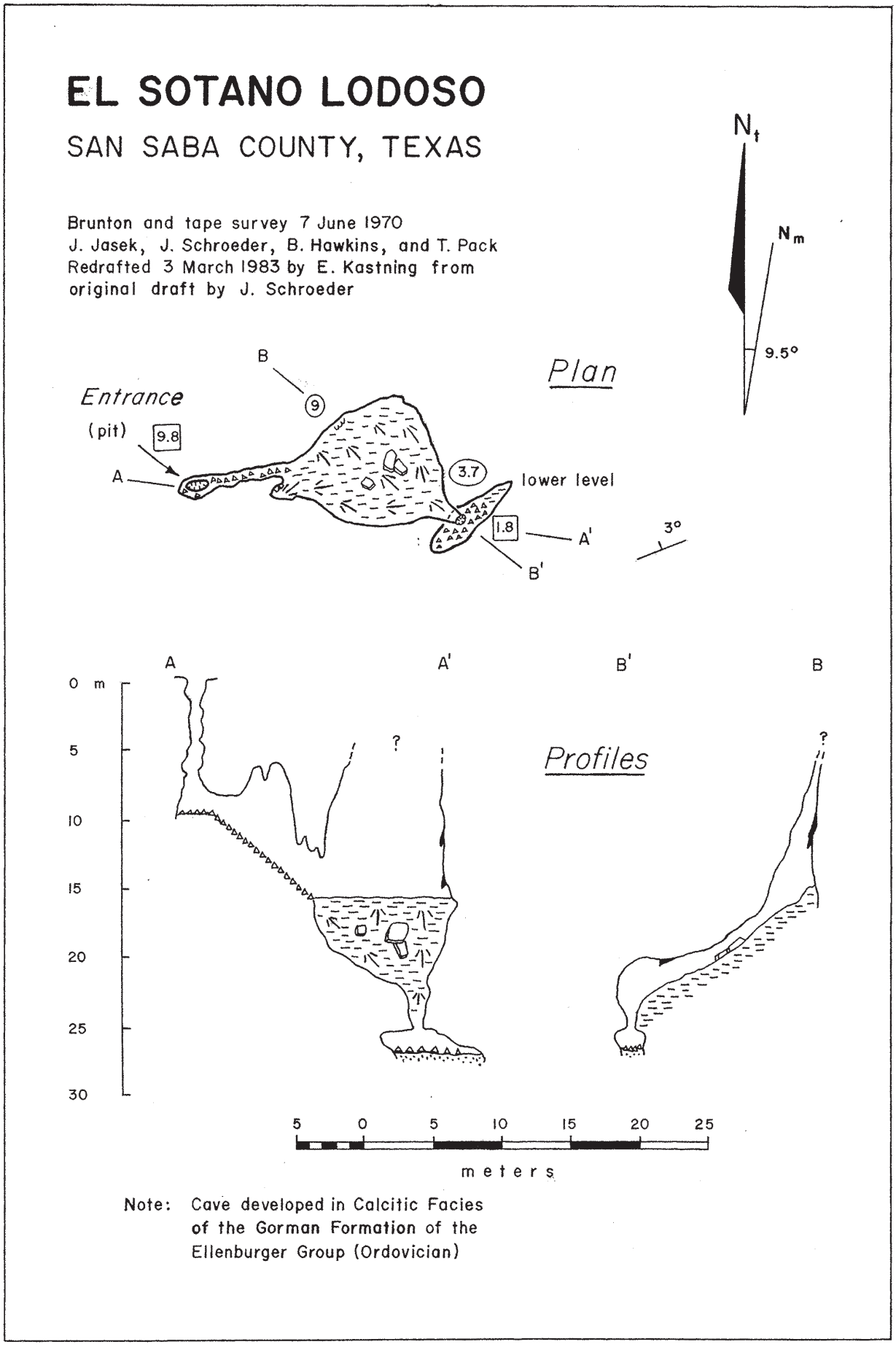


Figure 76: Map of El Sotano Lodoso. Modified from Reddell (1973, p. 103).

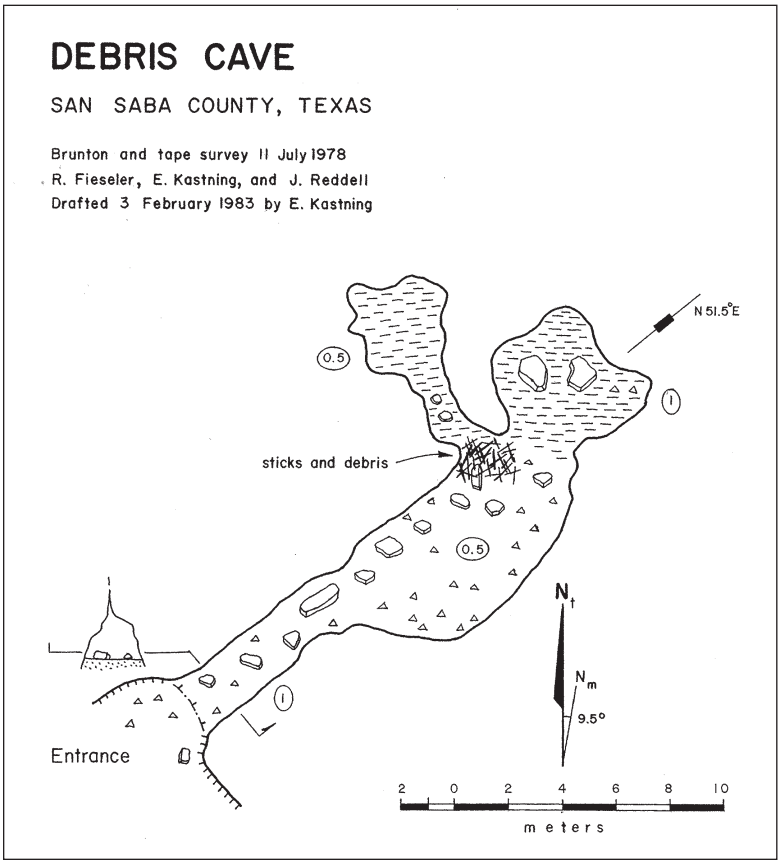


Figure 78: Map of Debris Cave.

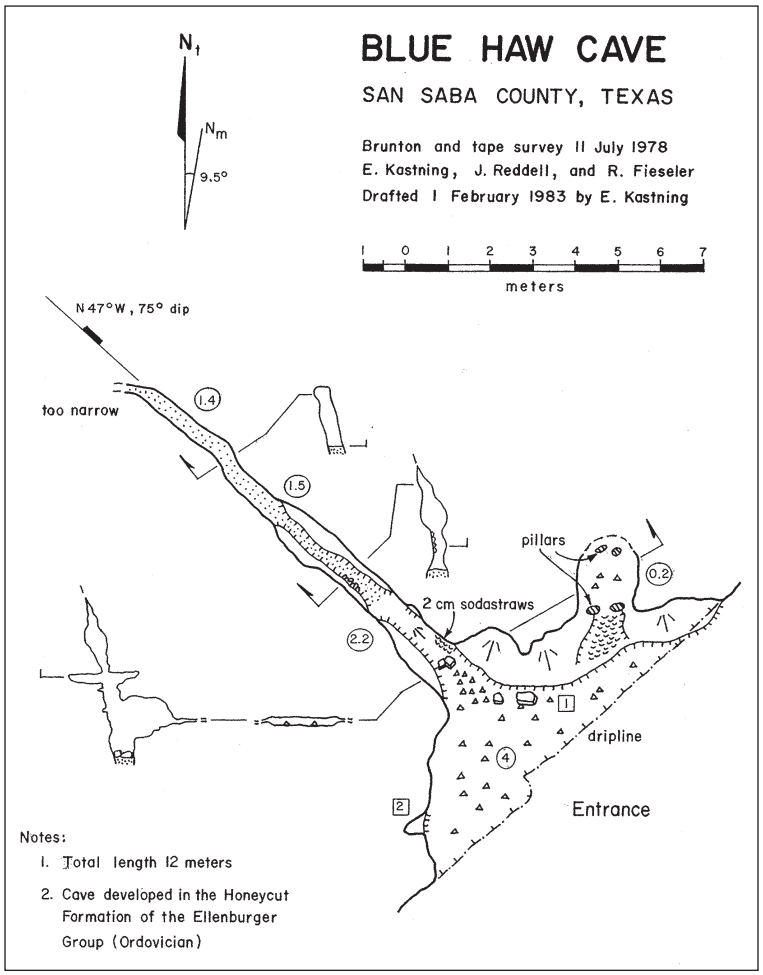


Figure 80: Map of Blue Haw Cave.

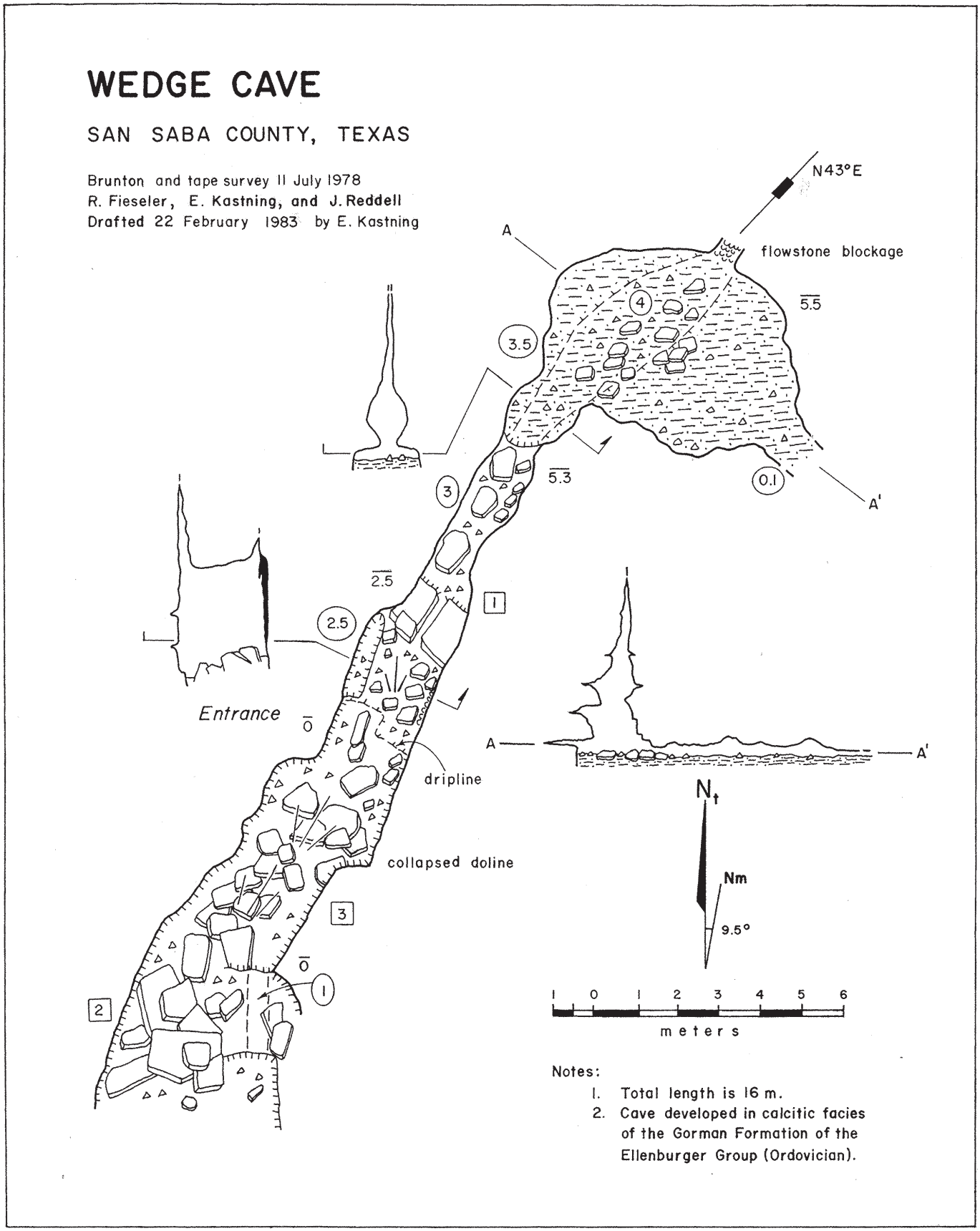


Figure 79: Map of Wedge Cave.

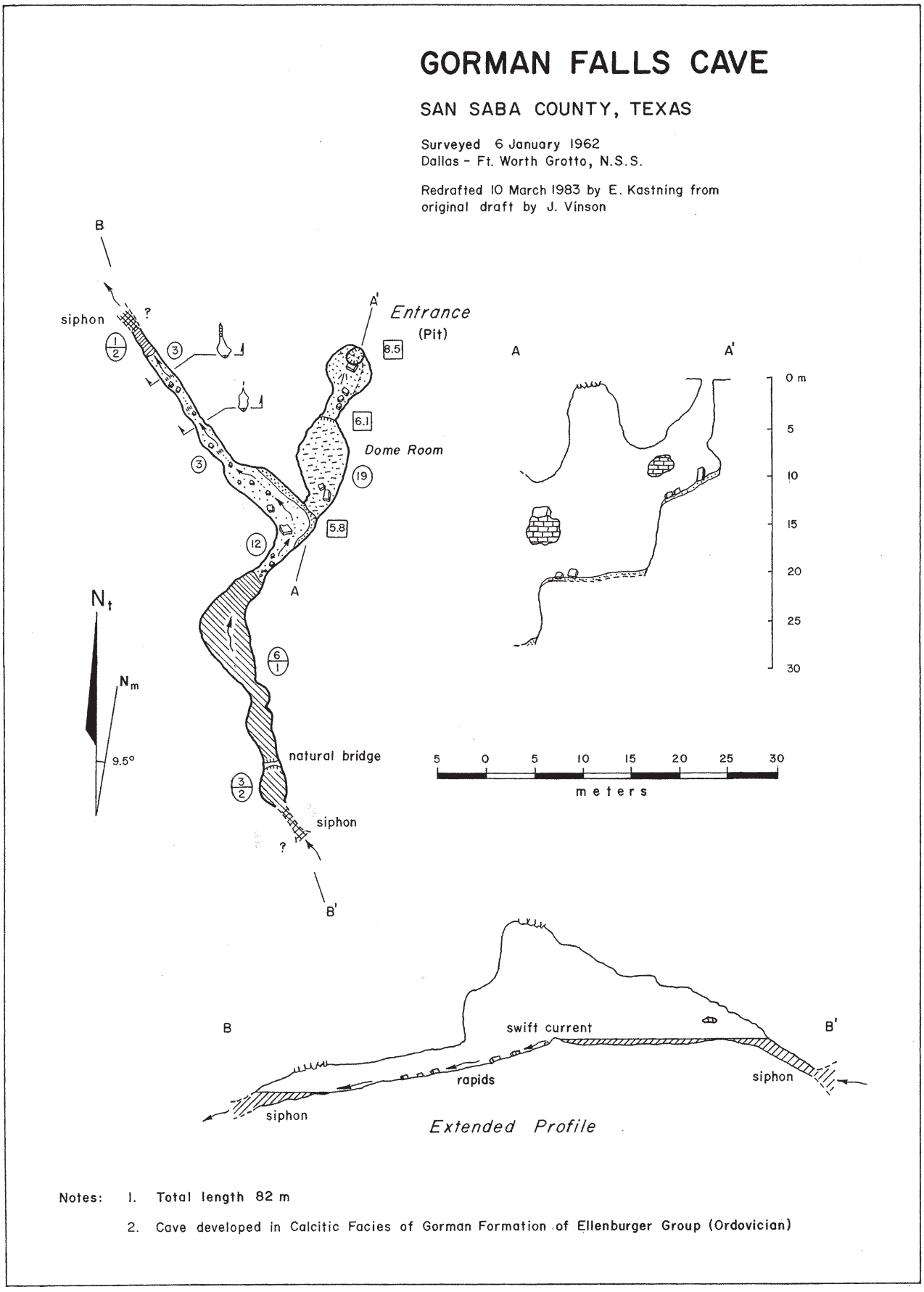


Figure 81: Map of Gorman Falls Cave. Modified from Reddell (1973, p. 61).

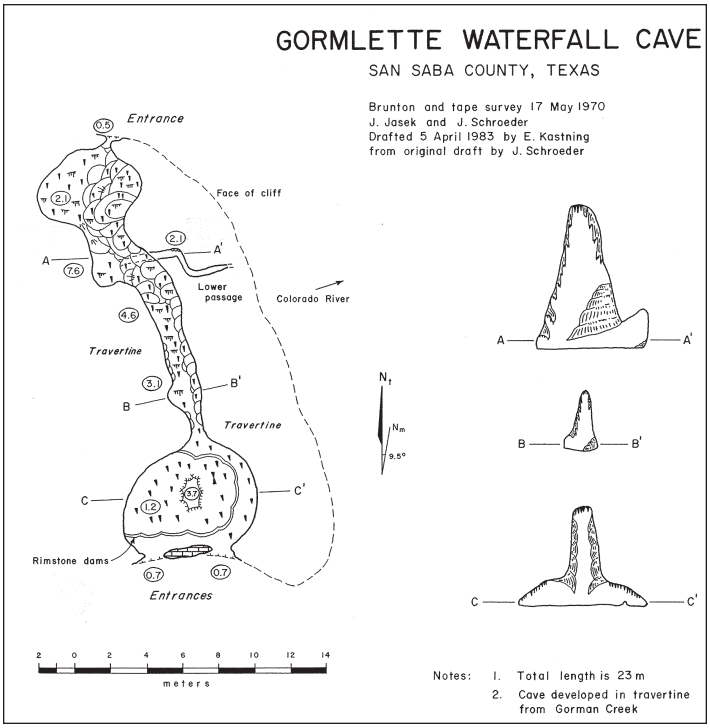


Figure 82: Map of Gormlette Waterfall Cave. Modified from Reddell (1973, p. 63).

Slumped Rock, Pokeweed, Three Entrance, Gourd, and Snail Fissures; Dagger Cave; Covered Crawlway Crevice; and Crack Pot). They are located on a topographic bench 120 to 1040 m northwest of Gorman Falls (Figures 51 and 73). The caves are vertically oriented fissures or crevices, developed along one or more fractures (Figures 83 and 84) and vary in length between 1 and 24 m and in depth between 2 and 11 m (Table 10; Figures 73, 74, and 85). The average length and depth are 11.8 m and 5.6 m respectively. Widths of the caves lie between 0.2 and 2 m and are generally uniform with depth. A few caves have one or two constricted entrances (e.g. Blacktail Pit, Figure 85P), but most are open to the surface along much of their length, except where they have become roofed by slumped or slid blocks of limestone (e.g. Meander Tube Fissure [Figures 85B and 86], Slumped Rock Fissure, Pokeweed Fissure, Three Entrance Fissure, Dagger Cave, and Covered Crawlway Crevice [Figures 85I,J,K,L,N, and 87]).

The caves are formed in the calcitic facies of the Gorman Formation and are floored by lithic fragments and surficial debris. Meteoric water freely infiltrates through these deposits to lower depths, but the caves remain dry throughout much of the year. Speleothems are rare or absent. Local baselevel for the caves is at the Colorado River to the east and 53 to 78 m lower in elevation.

Speleogenesis

The foregoing descriptions reveal that caves of the Bend vicinity are diverse in morphology, and that their genetic histories may vary considerably despite the small size of the area. The stratigraphic, lithologic and structural settings of the caves are similar; however, incision of the river, baselevels, topographic evolution, and the hydrogeologic setting have modified speleogenesis in different ways. Accordingly, the following discussion first addresses the control of lithology on speleogenesis, and then discusses other factors responsible for differences among caves. In this analysis caves are grouped morphologically as (1) major,

Figure 83: Offset Fissure, McLarrin Fissure Karst. Note separation of blocks at left along bedding planes. Depth to bottom is 6 m. View is to the southwest.

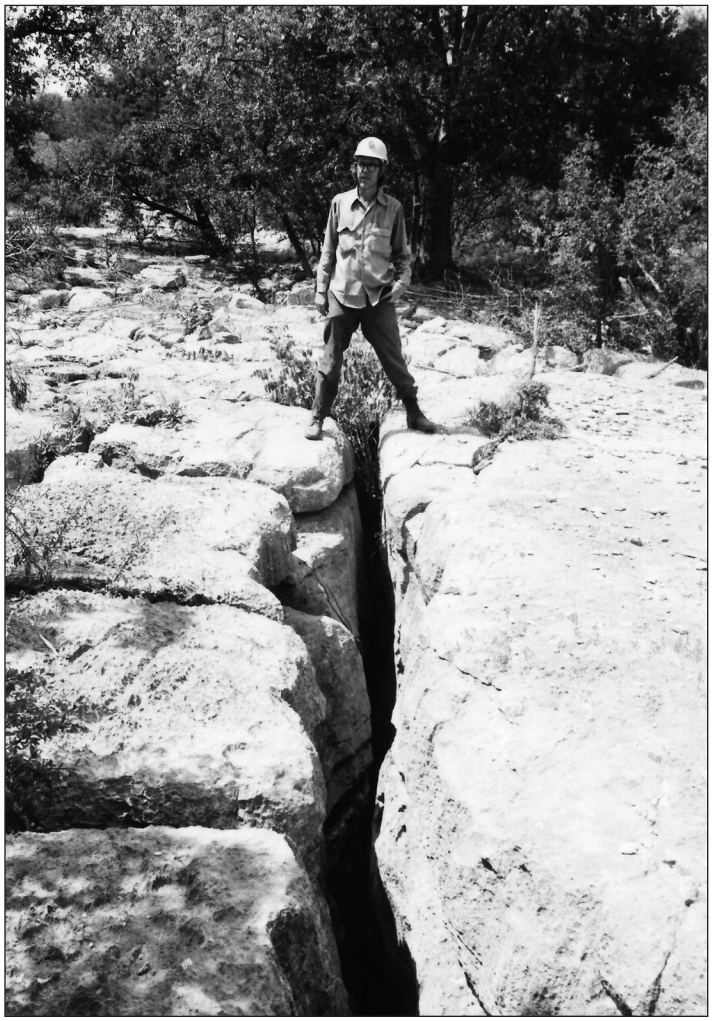
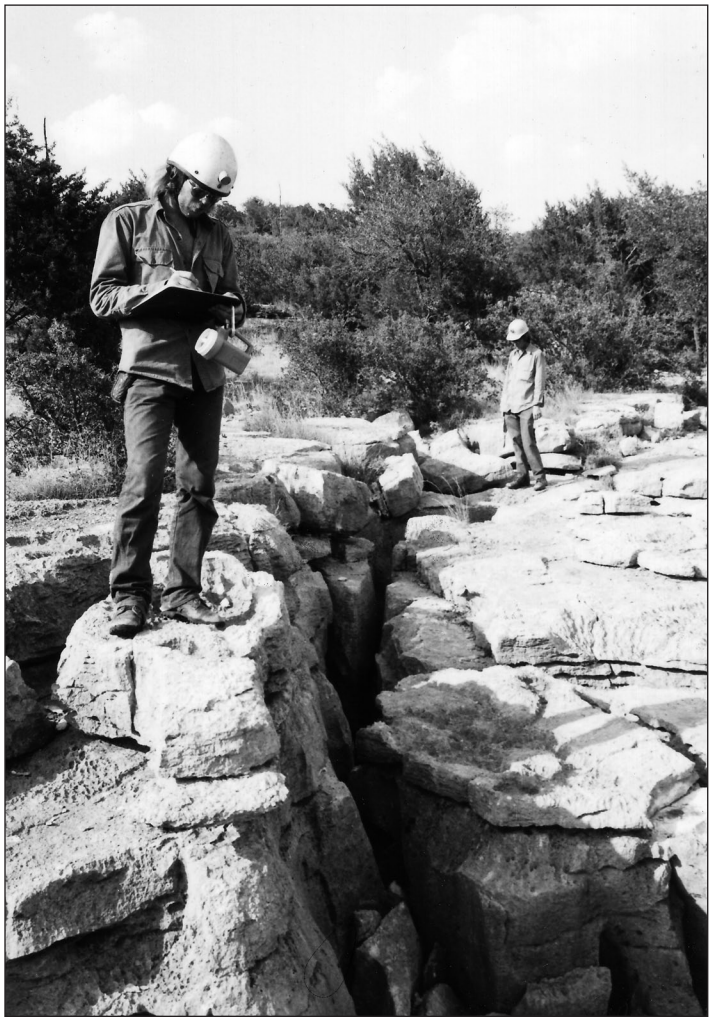


Figure 84: Crack Pot, McLarrin Fissure Karst. Note that the karstic pavement on right has been lowered by dissolution and erosion approximately 0.5 m more than that on the left. Dip is to the left. Depth to bottom is 6.3 m. View is to the northeast.



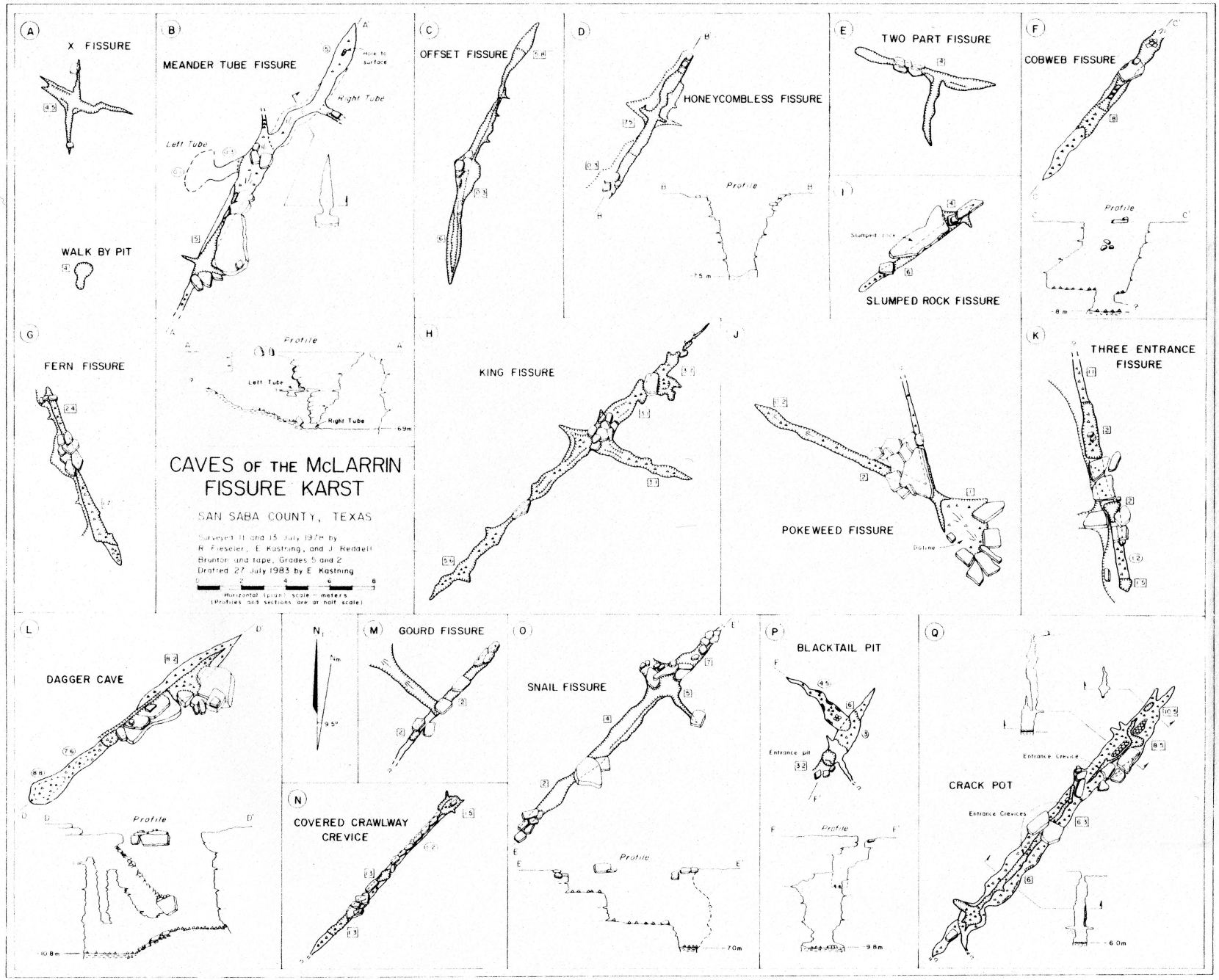


Figure 85: Maps of caves of the Mc Larrin Fissure Karst. Map of Daggar Cave modified from Reddell (1973, p. 38). For positions of caves relative to one another see Figure 73.

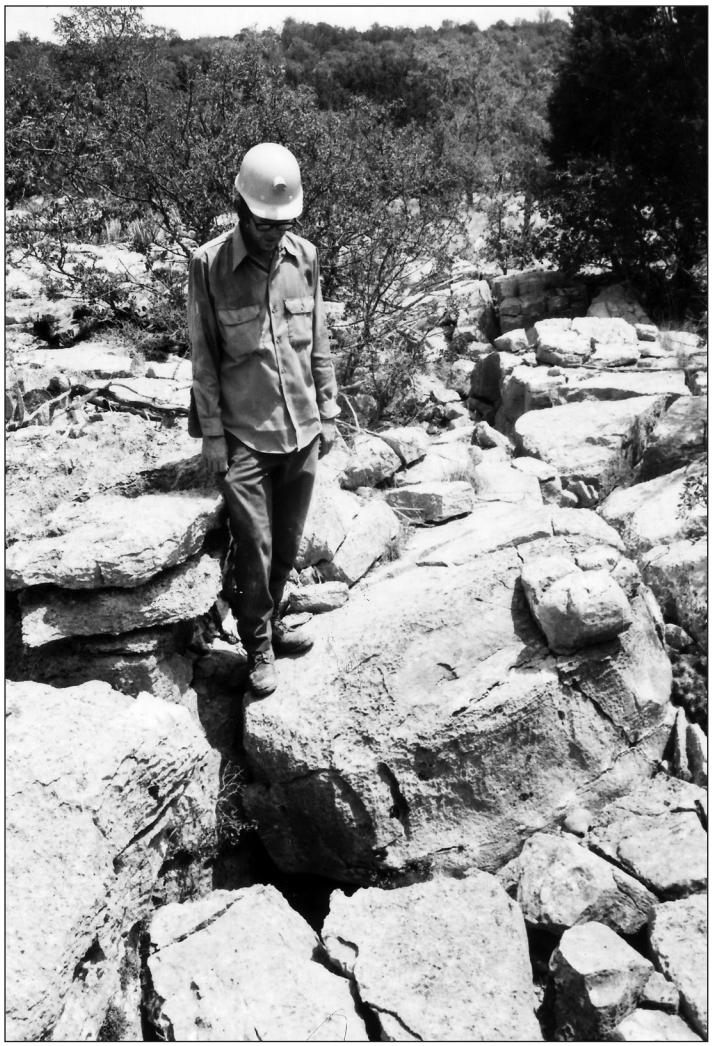


Figure 86: Meander Tube Fissure, McLarrin Fissure Karst. Note slumped blocks roofing the fissure and general disintegration and disruption of limestone-pavement. Depth to bottom of fissure is 5 m. View is to the northeast.

horizontal conduits, (2) short, horizontal caves, (3) deep, vertical caves, (4) fissure-like caves, or (5) caves in travertine.

Lithologic Control

Cloud and Barnes (1946a, p. 228-229) mention that the calcitic facies of the Gorman Formation in the Bend area is highly capable of transmitting groundwater, and that springs, small dolines, solutional openings, and caves are common in that unit. In contrast, they note that the Honeycut Formation is less cavernous, with considerably fewer springs and solutional features. These findings were confirmed by fieldwork and mapping of caves during the present study, and show that dissolution and development of caves have been enhanced along particular stratigraphic levels within the Gorman Formation.

The effects of lithology on excavation of caves is demonstrated by comparing positions of passages with the stratigraphic section (Figure 88). Detailed profiles of caves drawn in geologic sections through the area (Figures 74, 89, 90, and 91) show that solutional enlargement is greatest in the calcitic beds of the Gorman Formation and reduced in the dolomitic beds. Master conduits of large, cross-sectional dimensions, such as Gorman Cave, Clark's Branch Well Cave, and Lemons Ranch Cave, have formed in thick sequences of limestone (Figure 91); and the beds most favorable for speleogenesis are generally thin- to medium-bedded (3 to 35 cm), cleanly fractured, and aphanitic to medium-grained (Cloud and Barnes, 1946a, plate 14; see also Figures 54 and 88). Moreover, Gorman and Clark's Branch Well Caves, are concordant with the strata, are oriented down the dip, and are, in part, perched on dolomitic beds (Figure 53). The latter suggests that the lithology, composition, and texture of these units have made them relatively resistant to dissolution with respect to overlying beds of limestone.

Insight into the lithologic influences on speleogenesis can be gained by referring to the analysis of Ellenburger rocks by Goldich and Parmelee (1947). Although that study was performed on samples from the Warren Springs area of Llano County, their petrographic similarity to rocks of the Bend area (Figure 88) is sufficient for comparison.

Goldich and Parmelee (1947) determined the following properties of rocks of the Ellenburger Group (Figure 92):

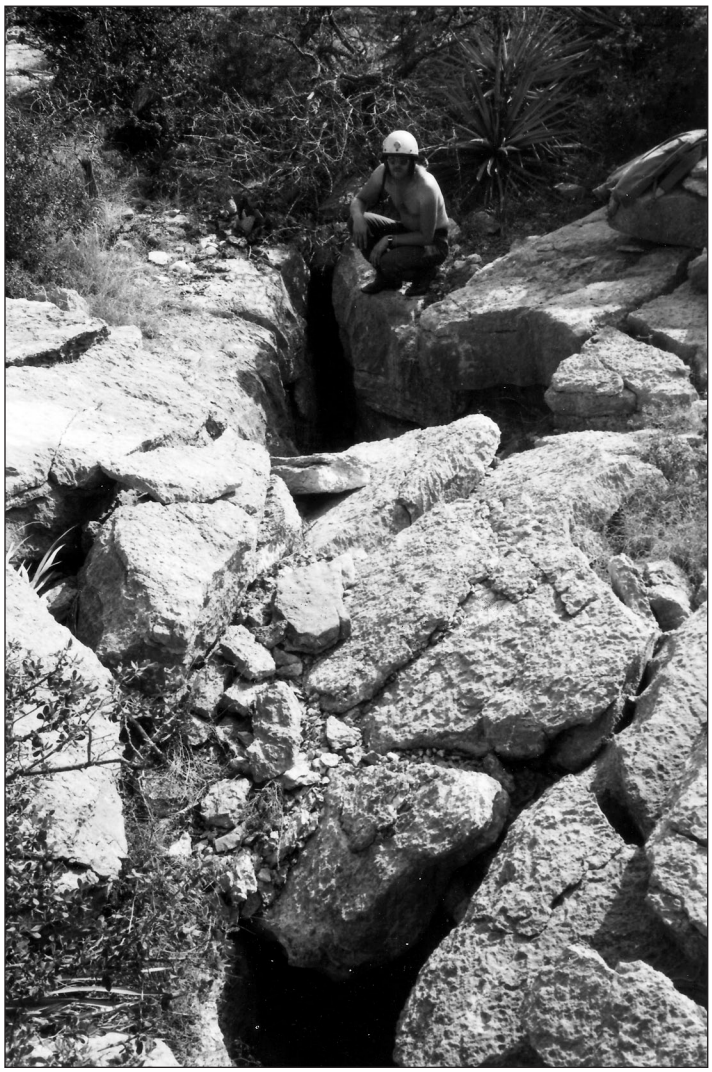
(1) Beds of limestone are very compact, exhibit low (0.1 to 0.7 percent), total porosity (volume of pore space divided by volume of the rock) and are low in content of insoluble material (0.4 to 6.5 percent).

(2) Beds of limestone are remarkably uniform in composition throughout the section (calcitic content ranges from 93.9 to 99.6 percent; dolomitic content ranges from 0.0 to 2.9 percent). In contrast, beds of dolostone are variable in composition (calcitic content ranges from 2.7 to 20.7 percent; dolomitic content ranges from 70.7 to 94.7 percent).

(3) The texture of beds of limestone throughout the section is generally uniform, ranging from aphanitic to medium-grained. However, the texture of beds of dolostone is more variable, with coarse-grained dolostone limited to the lower part of the Tanyard Formation, microgranular and very fine-grained dolostone restricted to the Gorman and Honeycut Formations, and fine- to medium-grained dolostone occurring throughout the Ellenburger section.

(4) The density of individual grains varies with composition and ranges from 2.707 grams/cm^3 (pure calcite) to 2.858 grams/cm^3 (pure dolomite) (Goldich and Parmelee, 1947, figure 3, p. 1995). Density of grains decreases with increasing grain-size for all beds of dolostone of the Ellenburger, with the exception of those of coarse-grained texture (Figure 92). This is caused by an increase in insolubles and calcite within the dolostone.

Figure 87: Dagger Cave, McLarrin Fissure Karst. Note slumped block which has subsequently been beveled by dissolution and erosion. Depth to bottom is 8.2 m. View is to the northeast. For general view of this area, see Figure 100.



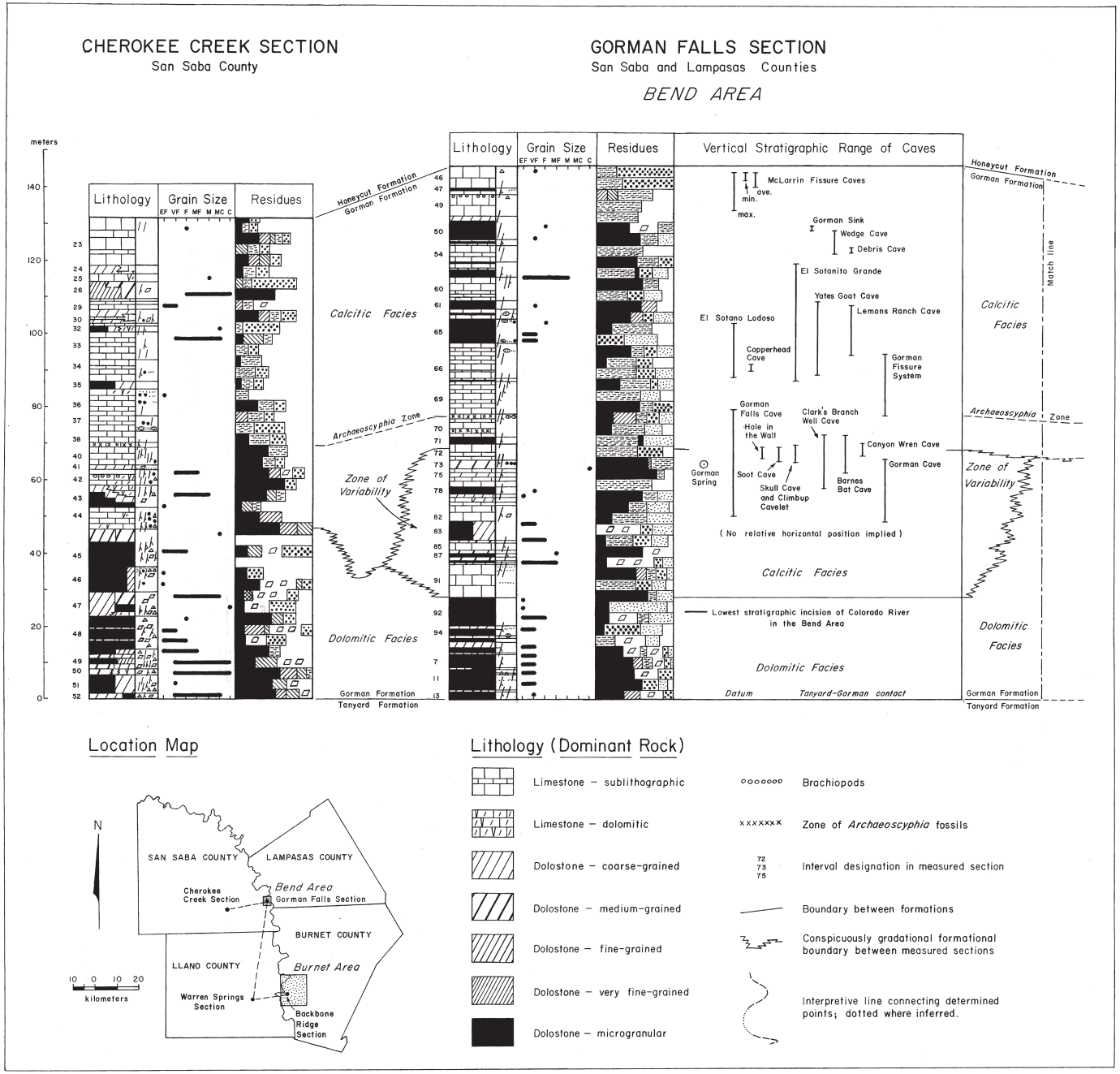
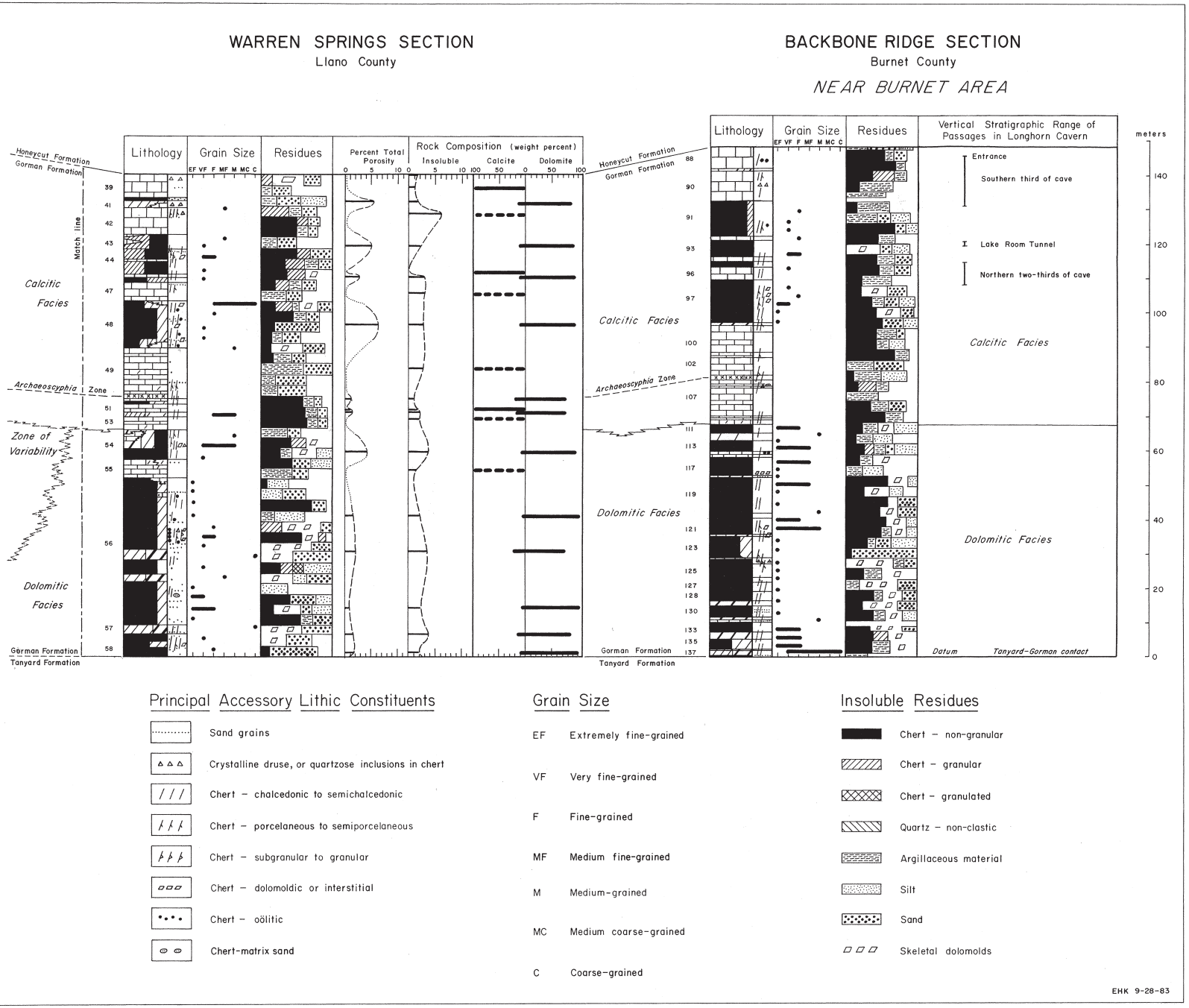


Figure 88 (here and below)**:** Correlation of rocks of the Ellenburger Group in the Llano Region. Columnar sections indicate dominant lithology, thickness of beds, total porosity, and insoluble residues. Stratigraphic positions of caves of the Bend and Burnet (Chapter 9) areas are shown. Compiled from Cloud and Barnes (1946a, plate 14). Hendricks (1952, plate 2), Goldich and Paralee (1947, Figure 8, p. 2016-2017), and investigations in the field. See also Figure 8.


WARREN SPRINGS SECTION
Llano County
BACKBONE RIDGE SECTION
Burnet County
NEAR BURNET AREA
Lithology
Grain Size
EF VF F MF M MC C
Residues
Percent Total Porosity
Rock Composition (weight percent)
Insoluble
Calcite
Dolomite
Vertical Stratigraphic Range of Passages in Longhorn Cavern
meters
Honeycut Formation
Gorman Formation
Match line
Calcitic Facies
Archaeoscyphia Zone
Zone of Variability
Dolomitic Facies
Tanyard Formation
Entrance
Southern third of cave
Lake Room Tunnel
Northern two-thirds of cave
Datum
Tanyard-Gorman contact
Principal Accessory Lithic Constituents
Sand grains
Crystalline druse, or quartzose inclusions in chert
Chert – chalcedonic to semichalcedonic
Chert – porcelaneous to semiporcelaneous
Chert – subgranular to granular
Chert – dolomoldic or interstitial
Chert – oölitic
Chert-matrix sand
Grain Size
EF Extremely fine-grained
VF Very fine-grained
F Fine-grained
MF Medium fine-grained
M Medium-grained
MC Medium coarse-grained
C Coarse-grained
Insoluble Residues
Chert – non-granular
Chert – granular
Chert – granulated
Quartz – non-clastic
Argillaceous material
Silt
Sand
Skeletal dolomolds
EHK 9-28-83

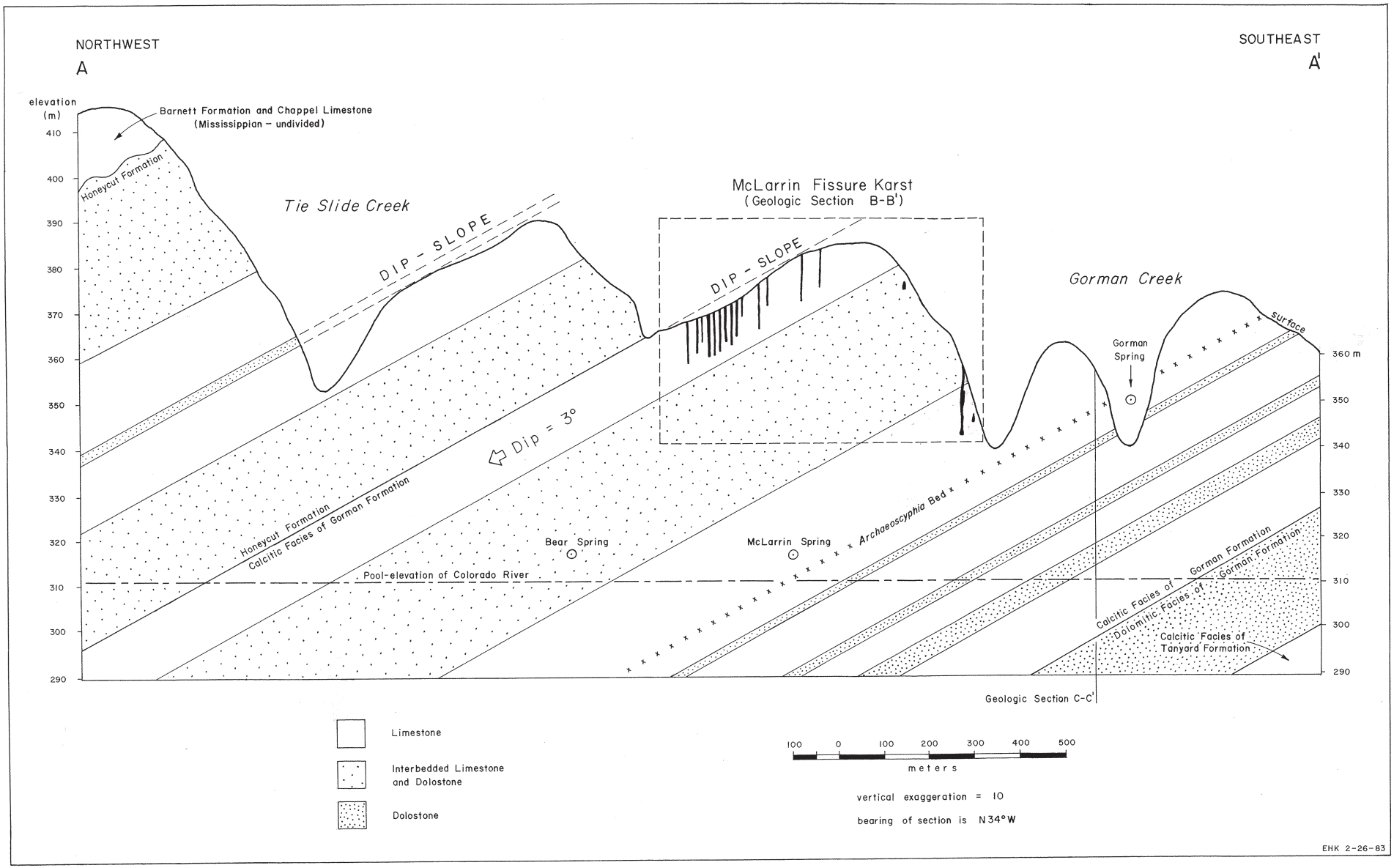


Figure 89: Geologic section A-A' of the northwestern part of the Bend area, showing stratigraphy and elevations of Colorado River and springs. Section enclosed by dashed line shown at expanded scale in Figure 74. For location, see Figures 51 and 69. Based on data in Cloud and Barnes (1946a, plate 9).

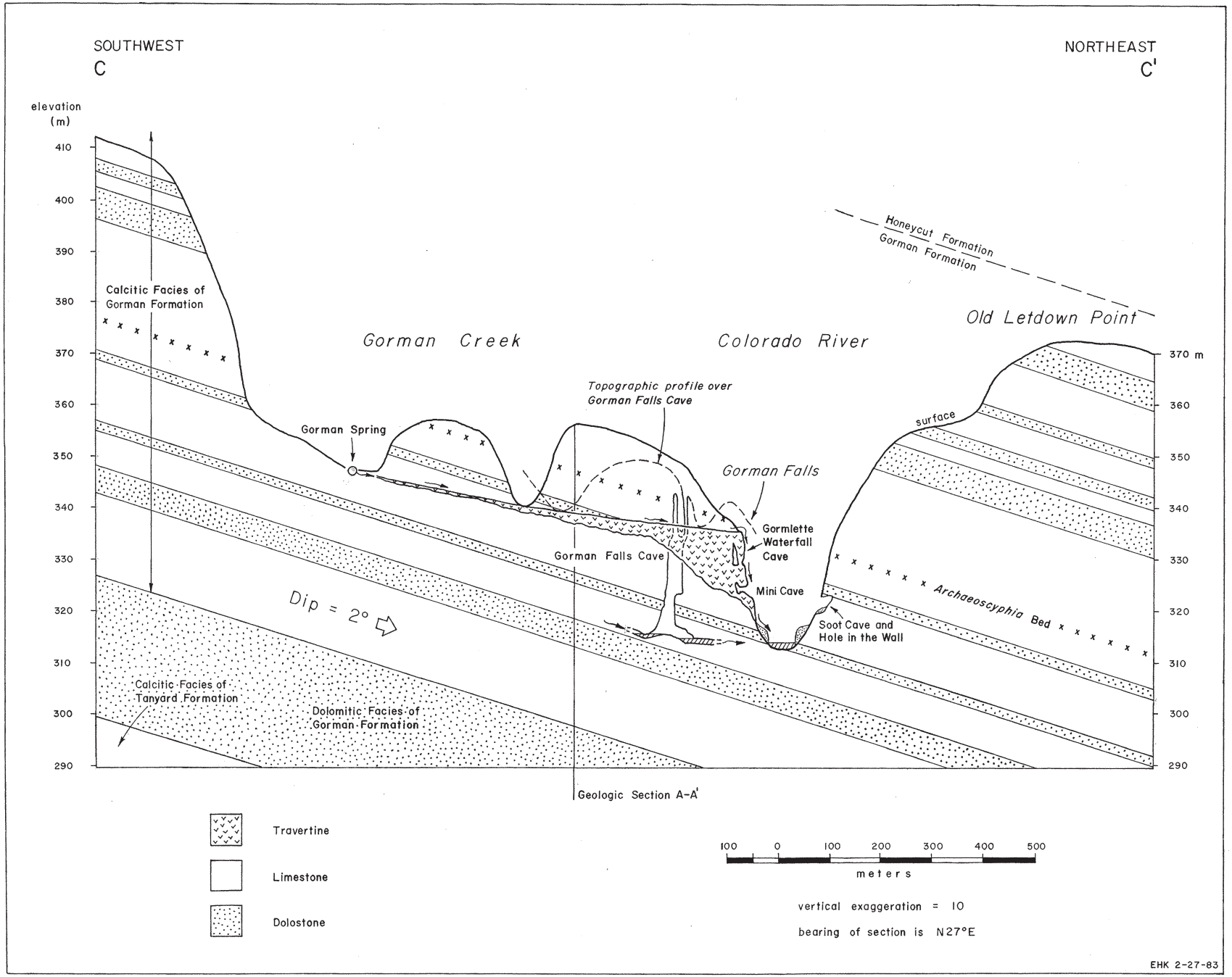


Figure 90: Geologic section C-C' of the northwestern part of the Bend area, showing stratgraphy, caves, Gorman spring and deposite of travertine. For location, see Figures 51 and 69. Based on data in Cloud and Barnes (1946a, plate 9).

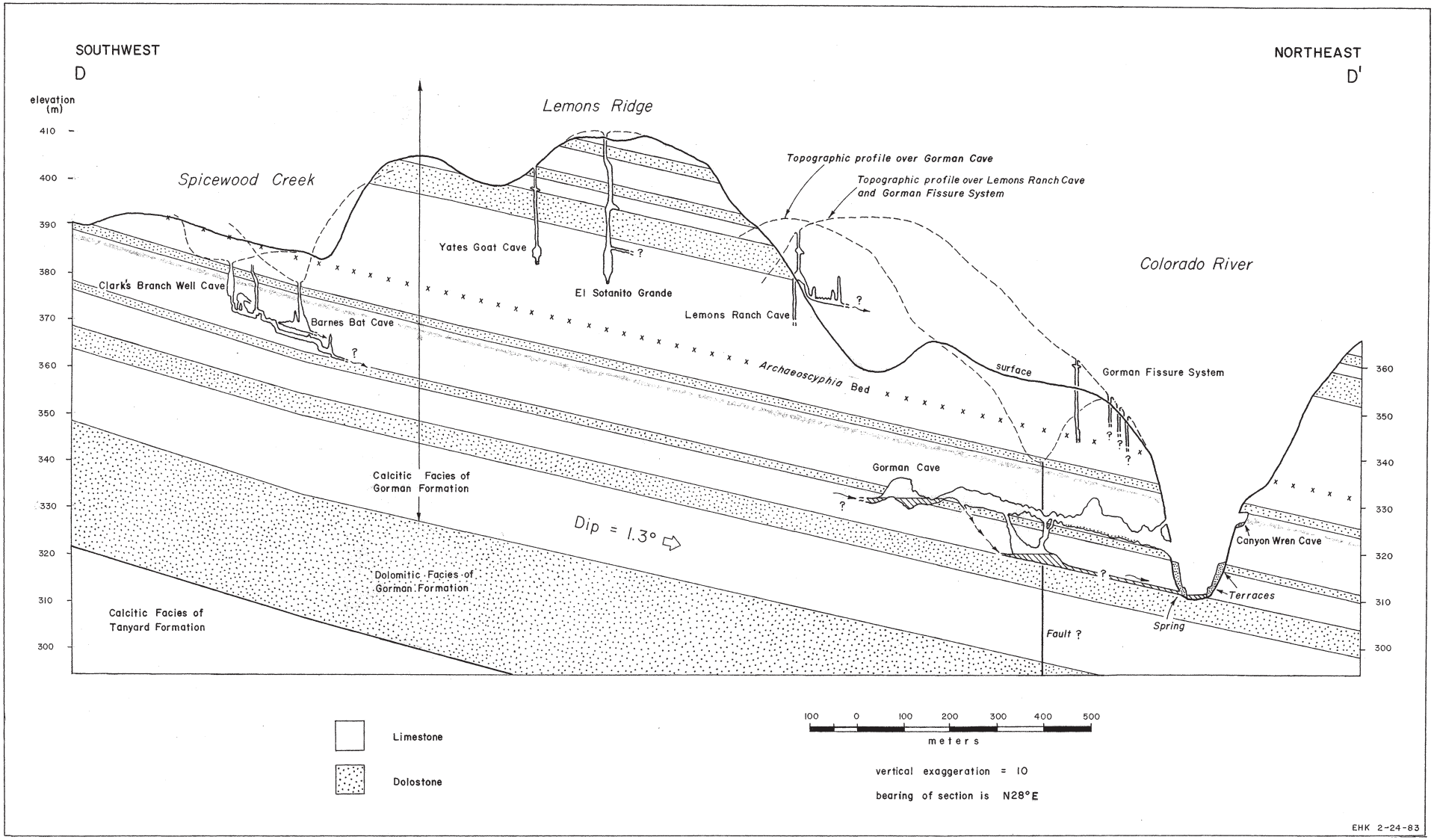


Figure 91: Geologic section D-D' of the southeastern part of the Bend area, showing stratigraphy, caves, and structure. For location, see Figures 51 and 64.

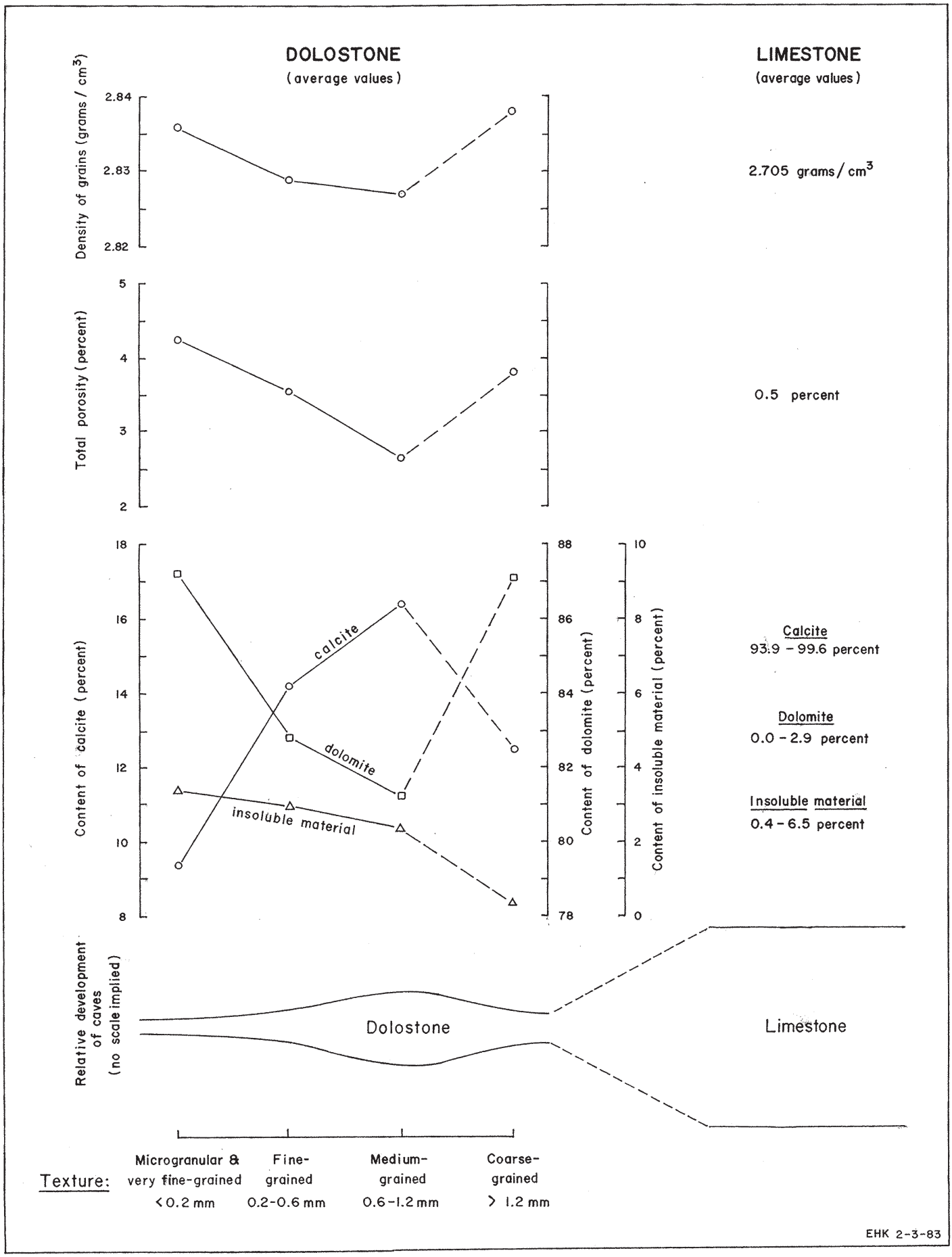


Figure 92: Variations in physical and chemical properties of caves of the Ellenburger Group and relation to development of caves Compare with Figure 88. Modified from Goldich and Parmalee (1947, figure 5, p. 2008 and table 8, p. 1998-1999).

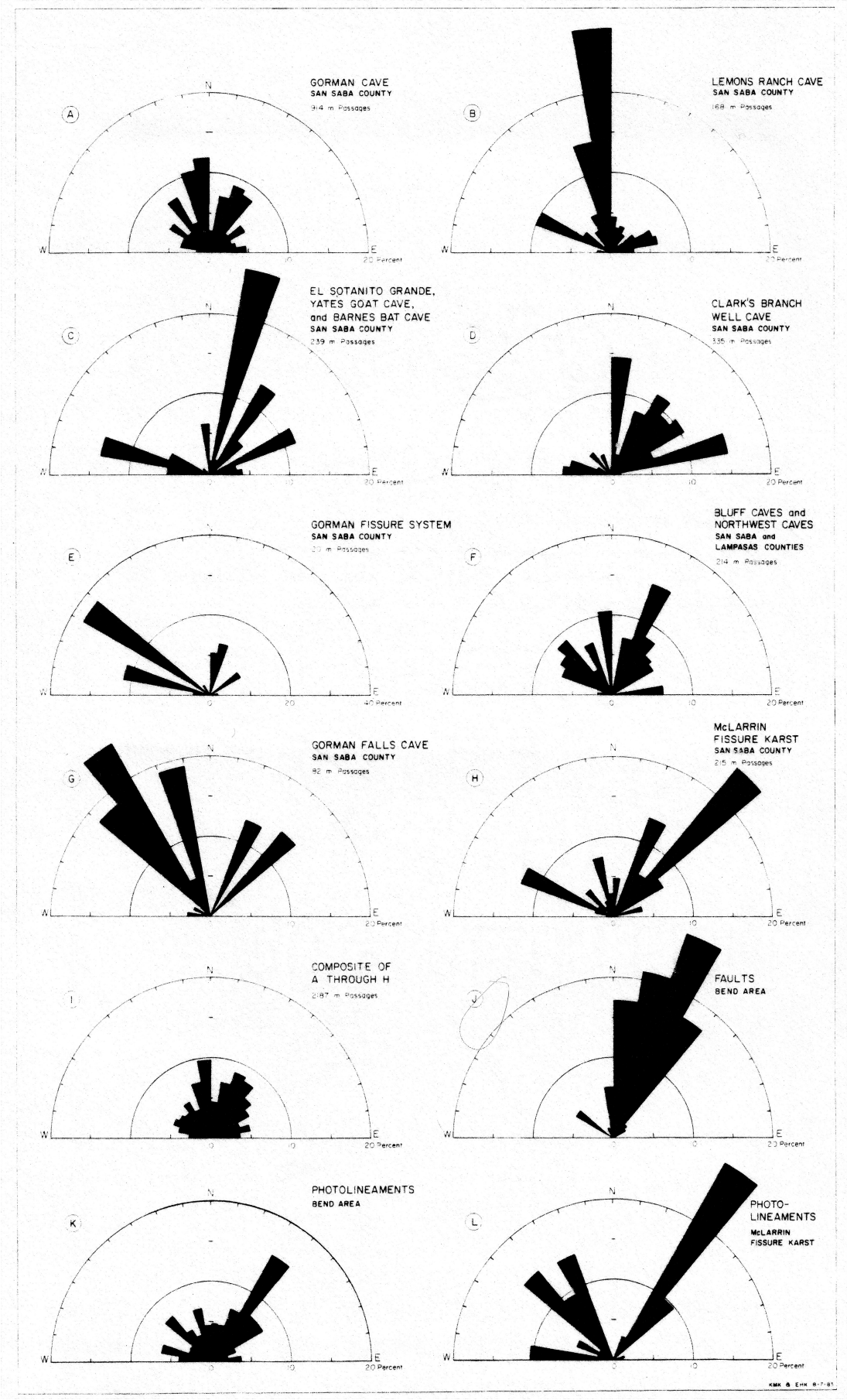


Figure 93: Rose-diagrams showing orientations of linear segments of passages in caves, mapped faults, and mapped photolineaments: (A) Gorman Cave, (B) Lemons Ranch Cave, (C) El Sotanito Grande, Yates Goat Cave, and Barnes Bat Cave, (D) Clark's Branch Well Cave, (E) Gorman Fissure Caves, (F) bluff-caves and northwest caves, (G) Gorman Falls Cave, (H) caves of the McLarrin Fissure Karst, (I) composite of all caves except caves in travertine, (J) faults in Bend area, (K) photolineaments in Bend area, (L) photolineaments in McLarrin Fissure Karst.

(5) Total porosity in the Gorman and Honeycut Formations alternates between relatively high values (for microgranular and very fine-grained dolostone) and low values (for dense, aphanitic to medium-grained limestone). In the beds of dolostone high porosity correlates well with high insoluble content (Figures 88 and 92). Microgranular dolostone exhibits the greatest porosity of all the Ellenburger rocks.

(6) Insolubles in the Ellenburger are largely chert and quartz (Figure 88). Lesser constituents include clay-minerals, feldspar, glauconite, iron-oxides, leucoxene, and traces of several others.

Stratigraphic positions of large caves (Figures 88 through 92) and the findings enumerated above indicate that the finest grained and most porous carbonates are the poorest cave-formers. This seeming paradox suggests that of all physical and chemical properties of the Ellenburger rocks, only content of calcite (purity of limestone) has exerted significant control on the stratigraphic positions of caves. Because the beds of limestone are exceedingly low in primary porosity, groundwater must have flowed through the rocks along zones of secondary porosity, such as fractures and collapsed areas (see discussion on structural control below).

The influence of bedding-plane partings and thickness of beds on excavation of caves has been minimal in comparison to the role of mineralic composition (calcite versus dolomite). The compact, dense, crystallinity of the rocks, and their relatively low porosities, have precluded any significant development of secondary permeability along bedding-planes. Compositional differences between calcitic and dolomitic limestones and their effect on dissolution and speleogenesis as observed in the Bend area are consistent with findings in other karstic regions (Sweeting, 1968; Thrailkill, 1977; Rauch and White, 1970; Rauch, 1972).

Structural Control

Visual inspection of maps of caves of the Bend area discloses that most passages consist of linear segments, and suggests that orientations of passages are significantly controlled by fractures. Furthermore, comparisons of orientations of caves with the attitude of bedrock indicate that many caves are developed parallel or subparallel to the dip. Detailed analysis of trends of fractures, photolineaments, and geologic cross-sections confirms these relationships.

Relationship to Attitude of Beds. Master conduits, such as those in Gorman, Lemons Ranch, Clark's Branch Well, and Barnes Bat Caves, are found only in the large fault-block in the southeastern part of the Bend area (represented by Figure 64). Conduits in this block are generally aligned in the direction of the north-northeastern dip toward the Colorado River (Figure 91). However, master conduits are conspicuously absent in the fault-block to the northwest (Figure 69), where the rocks dip to the northwest and parallel to the Colorado River (with the exception of Gorman Falls Cave near the river). Large caves are apparently best developed in areas where groundwater, confined to beds of limestone of high solubility between less soluble, dolomitic beds (see previous section), has been directly and efficiently conveyed down the dip to springs at baselevel along the river.

Fissure-like caves of the McLarrin Fissure Karst are limited in vertical extent to the 13 m-thick uppermost bed of limestone of the Gorman Formation. The erosional surface over many of these caves is a dip-slope developed just below the contact between the Honeycut and Gorman Formations (Figure 74). The fissures are blocked along their bottoms by sediment and surficial debris and presumably do not extend into the dolostone beneath. Drainage of the fissures is to the northwest, along solutionally enlarged bedding-plane partings (apparently formed during erosional unloading), and eventually along fractures striking northeast to small springs on the Colorado River.

Relationship to Fractures. The Bend area is cut diagonally by two major faults (Figure 51) which trend N 0°-40° E (Figure 93J) and divide the cavernous area into two major parts. In the west the Ellenburger strata strike N 60°-70° E and dip 2° to 3° to the northwest. In the east the strata strike N 75°-85° E and dip 1° to 2° to the northwest.

Lineaments were mapped from 1:20,000-scale, black-and-white, aerial photographs for the entire Bend area (Figure 94; see Appendix E for evaluation of the use of remotely sensed imagery for mapping lineaments and fractures). Dominant orientations are N 20°-60° E, N 10°-20° W, N 40°-50° W, and N 70°-80° W (Figure 93K). Many lineaments strike parallel to major faults, but others are apparently related to conjugate fractures formed during tensional stress at the time of faulting. Mapped lineaments are clustered in parts of the area (e.g. in the vicinity of the McLarrin Fissure Karst, along Tie Slide Creek, and near Gorman Cave), and they are more dispersed in the southern third of the area. Dense distribution of lineaments in the northern part of the area may be related to expansion of fractures in the deeply dissected areas along the Colorado River where unloading and release of confining pressures have been greatest. Moreover, in places, lineaments appear to be concentrated on outcrops of limestone, whereas fewer lineaments are observed on outcrops of dolostone, suggesting that limestone is more brittle; compare Figures 51, 89, and 94). A few lineaments radiate from zones of collapse and may be related to weaknesses induced by karstification during the Paleozoic.

Orientations of caves agree well with those of mapped lineaments (Figure 93, especially I and K). In the area west of the major faults lineaments dominantly trend N 30°-50° E and N 20°-50° W (Figure 93L), corresponding closely to orientations of passages in Gorman Falls Cave (Figure 93G), the McLarrin Fissure Karst (Figure 93H), and many of the Bluff and Northwest caves (Figure 93F). In the area east of the faults passages have strong north as well as northeastern or northwestern components (A through E of Figure 93). Here caves have been selectively excavated along fractures that follow the general northward dip, reflecting former flowpaths of groundwater along the steepest and most efficient, hydraulic gradients.

The role of faulting on speleogenesis in the Bend area is threefold. Firstly, faulting has caused disparate dips between the northwestern and southeastern cavernous areas, allowing extensive development of caves along the dip in the latter, whereas precluding such in the former. Secondly, fractures associated with and parallel to faults have provided some of the most significant avenues for flow of groundwater and dissolutional enlargement (compare I, J, and K of Figure 93). Thirdly, the displacements along the two major fault-zones have been large enough to compartmentalize the flow of groundwater toward the Colorado River. Circulation in the southwestern fault-block that includes Gorman and other caves may have been contained by boundary-faults as it flowed northward to the river.

Gorman Cave passes 11 m beneath the floor of a small valley about halfway along the length of the cave (Figure 64). The rock in the cave is severely shattered at this location, with breakdown littering and nearly blocking the passage (Figure 52). Some blocks of breakdown consist of breccia cemented with calcite. The valley above forms a lineation in the topography, that is coincident with a photolineament traceable for 3.7 km across the Bend area (Figure 94). Cloud and Barnes (1946a, plate 9) show two, short (130 m) faults just north of the cave on the other side of the Colorado River, one of which coincides with this lineament. I interpret the fault to extend to the southwest where it intersects Gorman Cave at the zone of collapse (Figures 52, 64, and 91), and I have added it to the geologic map (Figure 51). It is interesting that the position of the fault marks the downstream, explorable limit of the lower-level passage in the cave, and that 150 m south of the fault-zone the stream is perched at a higher level. Moreover, passages just south of the cave bifurcate and form small maze-like networks at the Swiss Cheese and near the Crack (Figure 52). These complex areas may have been formed by backflooding behind constrictions that formed from collapse at the fault. Development of routes of diversion and mazes in response to catastrophic floods has occurred in other areas of the Edwards Plateau (see Chapters 6, 7, 10, 11, and 12). Discussions of the hydrodynamic mechanisms of flooding with regard to speleogenesis are presented by Palmer (1971, 1972, 1975)

Another intensely fractured zone in Gorman Cave occurs 6 m north of the Big Room (Figure 52). Although breakdown is absent at this location, walls of passages consist of breccia cemented with calcite. Immediately upstream the large dimensions of the Big Room and the presence of joint-spurs, including the Mouse Room, suggests that the brecciated zone served as a constriction behind which floodwaters ponded and enlarged the passage. Whether the fractured zone represents another fault or perhaps a collapsed structure similar to those mapped by Cloud and Barnes (1946a) could not be ascertained. In any case, this well-cemented breccia was apparently more resistant to solution than the undisturbed limestone either upstream or downstream.

Synergetic Effects of Lithology, Structure, and Hydrogeology. As discussed above, speleogenesis is confined to beds of highly pure limestone having very low primary porosity. Without development of secondary porosity virtually no circulation of groundwater nor dissolution would have taken place. Avenues through which groundwater could come in contact with soluble limestone were provided through fracturing. Fractures are spaced as little as a few meters apart in places where the frequency of fractures has been relatively high owing to faulting, denudation, and brittle beds of limestone. Rates of circulation of groundwater and dissolutional enlargement of master conduits had been intensified along such zones. Caves are composed of linear segments and controlled by fractures because bedding-plane partings between the compact, dense carbonate beds remained closed, whereas vertical fractures were initially open to circulation of groundwater. Tensional stress in the Bend area during the Llano uplift created block-faults and open joints. Near the Colorado River joints may have expanded further from dissection of the landscape and removal of overburden.

Major solutional conduits develop along flowpaths that have the largest, initial

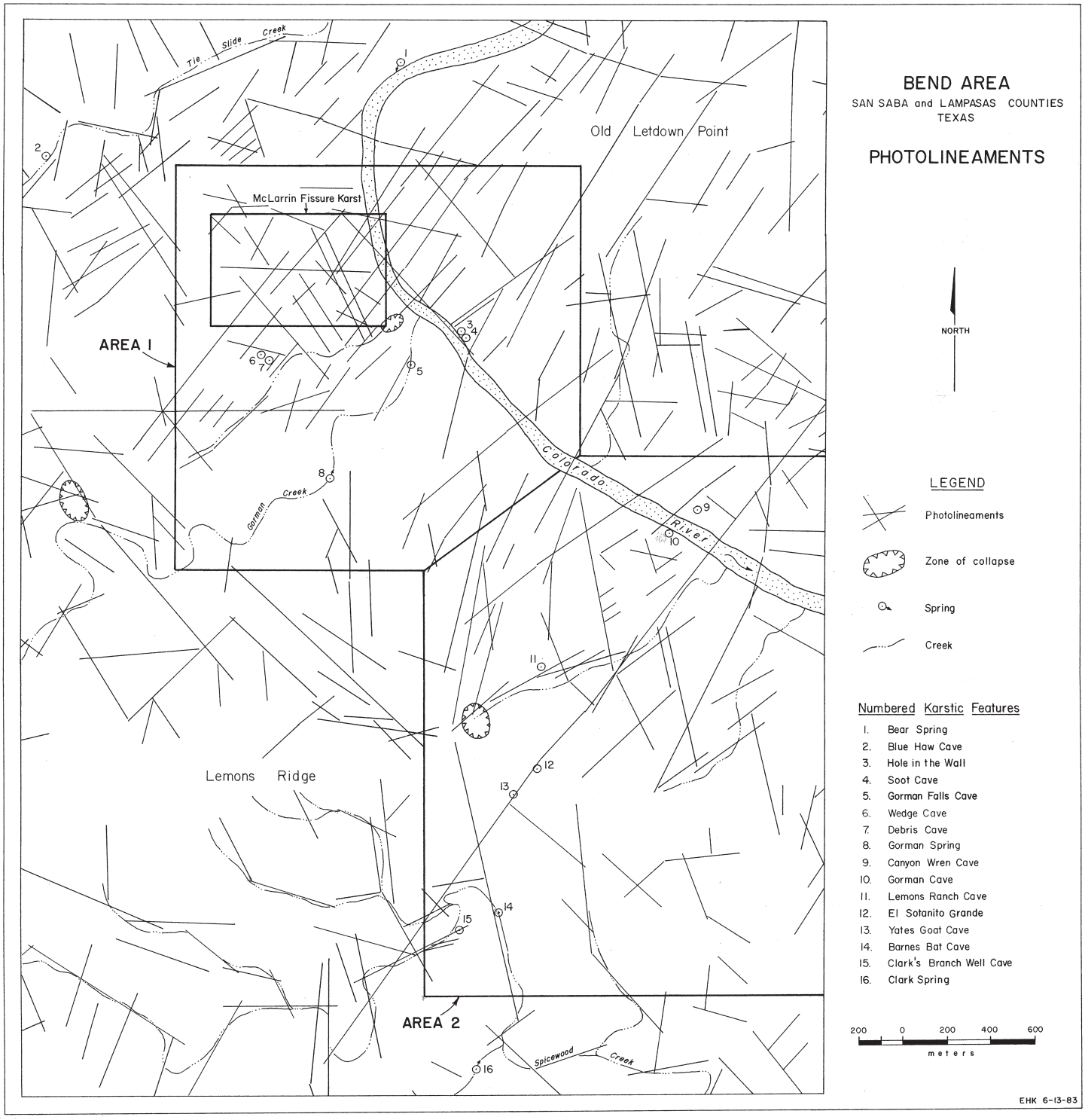


Figure 94: Photolineaments of the Bend area. Heavy boundaries represent areas of Figures 64, 69, and 73. Mapped from U.S. Department of Agriculture 1:20,000, black and white, panchromatic, stereo-pair imagery: image numbers BRH-1W-21 through 24 and BRH-1W-45 through 48, flown 1 November 1958. Area is the same as in Figure 51.

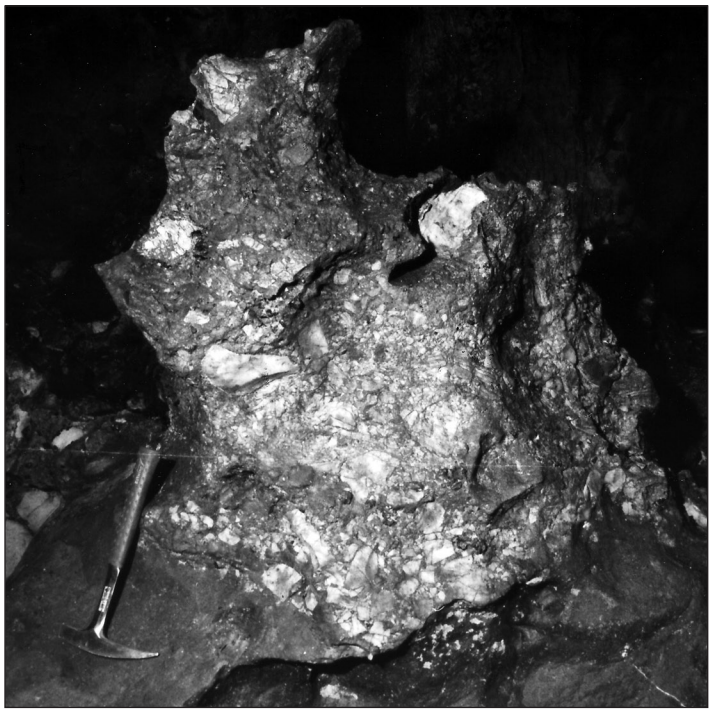


Figure 95: Pedestal in Gorman Cave, approximately midway between Separation Lake and the Squeeze. Top of pedestal consists of well-cemented breccia and has been scalloped. Note streamlined form, tapered at right, Flow occurred from left to right. Base of pedestal is bedrock on floor of cave (Gorman Formation). View is to the west. Hammer is 33 cm long.

cross-sectional area and steepest, hydraulic gradients. Consequently, the largest conduits in the Bend area are positioned (1) along the dip, which provides the steepest hydraulic gradients toward the Colorado River, and (2) along north- to northeast-trending fractures, which initially were the most open.

Locally, some flowpaths may have developed along highly fractured and brecciated zones associated with collapsed structures or faults. Collapse that occurred during the Paleozoic may not be significant in this regard because collapsed dolines were later filled with younger carbonate deposits, and any porosity resulting from collapse has subsequently been removed by infilling, cementation, and compaction. However, collapse related to faulting and incision by valleys can either promote or impede the flow of groundwater (Kastning, 1977c). In Gorman Cave flow has been transmitted a short distance through two brecciated and collapsed areas, but these zones, in turn, have formed constrictions to flow in the master conduits, leading to backflooding, locally increased hydraulic gradients, and greater turbulence of flow.

The hydraulic effect of the constriction between the Funnel and Separation Lake in Gorman Cave is evident from sculpted features about 25 m downstream (north) of the fault (Figure 52). A 1.5 m-high pedestal composed of a well-cemented breccia at the top and bedrock at the base (Figure 95) is positioned in the center of the passage which is 3 m wide at this point. The surface of this feature is scalloped and cavitated, and in top-view, the brecciated component of the pedestal is streamlined, with a blunt surface facing upstream (south) and tapered edge facing downstream (north). This morphology indicates flow to the north and the presence of turbulence, a requisite for scalloping. Approximately 3 m north of the pedestal are similarly sculpted forms in the ceiling of the passage (Figure 96). These sculpted indicators of flow suggest: (1) The passage was constricted from this point to the Swiss Cheese by collapse and by brecciation created during faulting in this vicinity; (2) The well-cemented breccia and zone of breakdown resisted dissolution as enlargement of conduits progressed elsewhere along the cave; (3) Velocities of flow through the constriction were higher than the average velocity through the cave and sufficiently fast enough to cause turbulence in the constriction and immediately downstream (see Palmer, 1972, for discussion of hydrodynamics at constrictions); (4) The pedestal and sculpted ceilings and walls are relicts from periods of high discharge and are now 1 to 3 m above the present floor of the cave; (5) During periods of high discharge, then as today, water unable to be transmitted along the smaller, lower-level passage overflowed through the Funnel and similar holes into the main (upper) conduit.

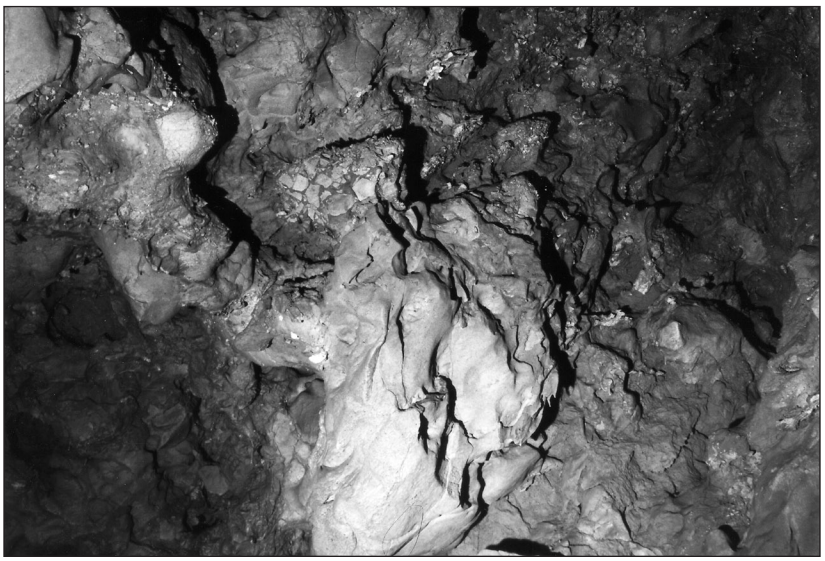
The interrelationship between lithology and structure may be summarized as follows: (1) Caves have developed preferentially within highly calcitic beds of limestone, with little development in units of dolostone; (2) Water was transmitted down the stratal dip exclusively along vertically oriented fractures; (3) Rates of dissolution were greatest where water was transmitted along open, tensional fractures in the beds of limestone and where hydraulic gradients were steepest toward the river; (4) As a result fracture-controlled caves, which appear angulate in plan view and are stratigraphically positioned along highly soluble beds, are dominant in the Bend area. This is in sharp contrast to caves of other areas of the Edwards Plateau that are controlled by bedding-plane partings or have developed in highly porous, carbonate beds. Typically, such caves exhibit a meandering or sinuous form, or consist of irregularly or vugularly shaped chambers.

Topographic Control

The baselevel toward which groundwater flowed during speleogenesis was the Colorado River. Incision of the Colorado began along the Balcones Escarpment during the Miocene, in response to regional uplift and Balcones faulting (Chapter 2). However, an erosional surface had developed over the interior of the Edwards Plateau and Llano Basin by the mid-Pleistocene. Evidence for this includes deposits (e.g. Uvalde gravel along the eastern and southern margins of the plateau) which were eroded and transported across the erosional surface at that time. The Colorado River had developed considerable sinuosity as it flowed across the erosional surface. This implies that this surface was then nearly horizontal. The Colorado began to incise the surface in the mid-Pleistocene as it tended to become graded to the new baselevel at the Balcones Escarpment. Moreover, the present Colorado River is underfit with respect to its valley, indicating that discharge during the late Pleistocene, during pluvial conditions, was considerably greater than it is today. Meanders that were formed prior to the mid-Pleistocene, including those in the Bend area, were subsequently preserved during entrenchment. These events occurred along many rivers of the plateau, as suggested by incised or abandoned fluvial deposits, meanders, and other relict features. Existing meanders of the pre-Miocene Colorado have been preserved in the Bend area during entrenchment.

It is difficult to ascertain any distinct correlation between levels in caves and stages of incision by rivers. Lithology and stratigraphic position appear to dominate as controls on horizontal development of caves (Figures 89, 90, and 91). However, the vertical extent of some deep caves is apparently related to the high relief created by incision. Low elevations of some caves with master conduits (e.g. Gorman, Clark's Branch Well, Barnes Bat, and Gorman Falls Caves) with respect to the present level of the river suggest that these caves are relatively young (perhaps as recent as the late Pleistocene). Support for this comes from the positions of springs along the river. Bear and McLarrin Springs lie approximately 6 m above

Figure 96: Solutional sculpture in ceiling of Gorman Cave, along east wall of passage approximately 3 m north of pedestal shown in Figure 95. Note streamlining of breccia and bedrock. View is to the east. Width of photograph includes 2 m of ceiling.



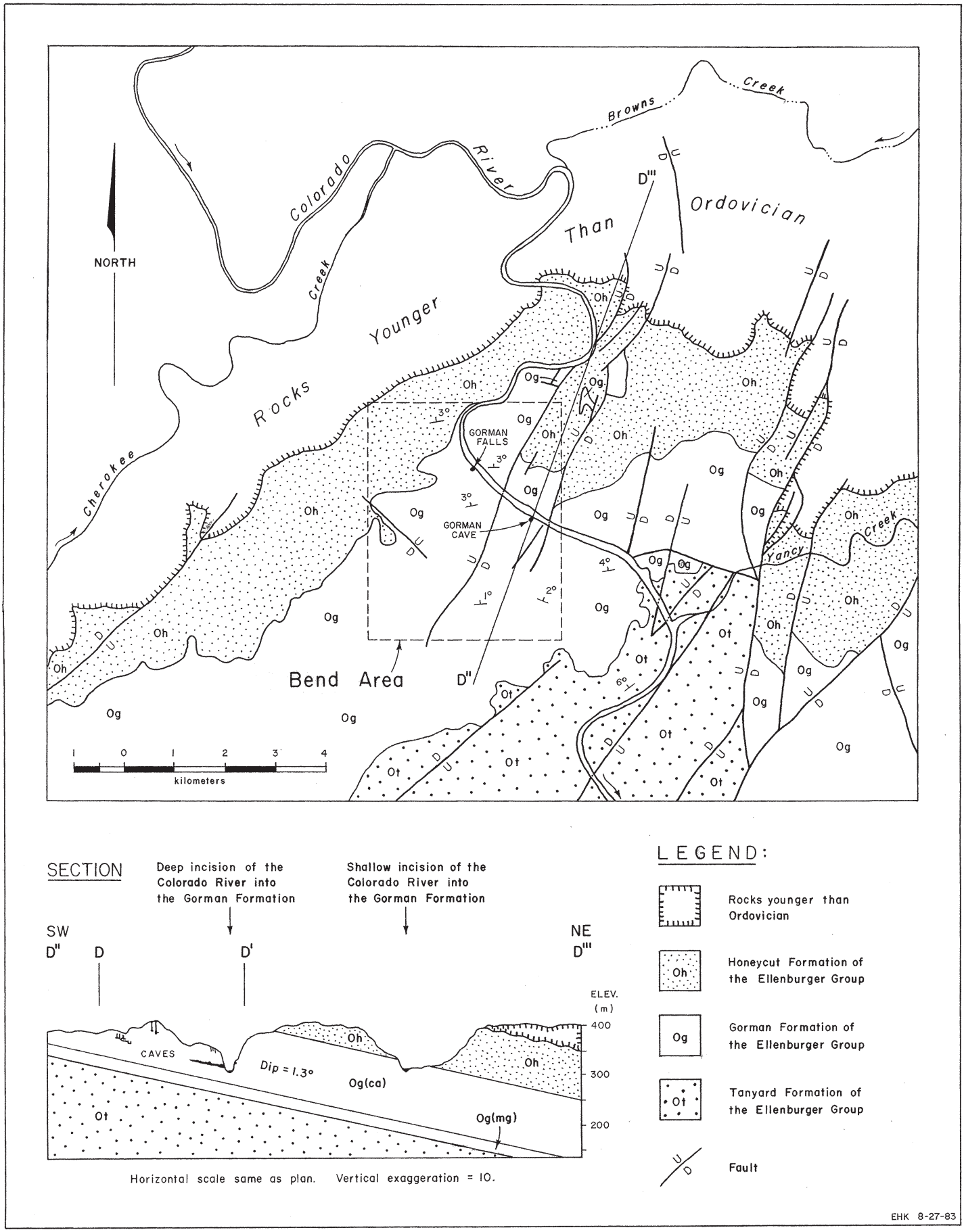


Figure 97: Geologic map of the northeastern Llano Uplift showing excursion of Colorado River across the Gorman Formation. Development of caves has been greatest along this reach where dips are to the north. Detailed section D-D' is shown in Figure 91. Modified from Cloud and Barnes (1946a, plate 9) and Barnes (1976a).

the level of the river (Figure 89) and may represent former points of discharge in the phreatic zone. The small spring along the edge of the river at Gorman Cave appears to discharge flow from the lower level of Gorman Cave, which is presently at grade with the river (Figure 91).

Gorman Cave is within the present flood-zone of the Colorado River. The river has been observed to rise to a level which allows water to back up into the cave (Beck, 1970). Evidence for enlargement of the cave by floodwater is scanty. Pockets in ceilings and enlarged joint-spurs occur in places (Figure 54). Coarse debris, consisting of sand, pebbles, and gravel has been carried into most of the cave (Figures 54 and 55), and in places such deposits are interbedded with clay and silt (Figure 56). Apparently Gorman Cave has been only moderately enlarged by floodwater from the Colorado River.

The Bend area contains the only extensive caves for some distance along the Colorado River. This can be explained by analyzing the relationship of incision of the river to the topography and regional dip. Topographic and structural conditions in the Bend area are similar to those in the Langtry area (Chapter 5); however, the carbonate rocks in the Bend area are much less vertically homogeneous than those at Langtry, and caves are thereby less vertically developed.

The Colorado River makes a sweeping curve to the west-southwest as it crosses the Ellenburger outcrop in the Bend area (Figure 97). At its maximum westward excursion it passes through the Gorman Formation, which represents the most soluble zone of the Ellenburger Group. Because beds dip to the north, especially in the cavernous, southeastern fault-block, dissection of the Gorman Formation is stratigraphically deepest along the meander, allowing a thicker zone for the development of caves, greater infiltration in the uplands, steeper hydraulic gradients, and augmented discharge from conduits to springs at baselevel (see cross-section in Figure 97).

Origin of Bluff-Caves

Caves along the Colorado River and Tie Slide Creek (Canyon Wren, Soot, Skull, Climbup, Blue Haw Caves and Hole in the Wall) were formed when the Colorado River and its tributaries were at a higher elevation. These small caves served as local points of discharge and later may have been enlarged by periodic flooding of the Colorado. The bluff-caves are older than those containing master conduits and are now relict features.

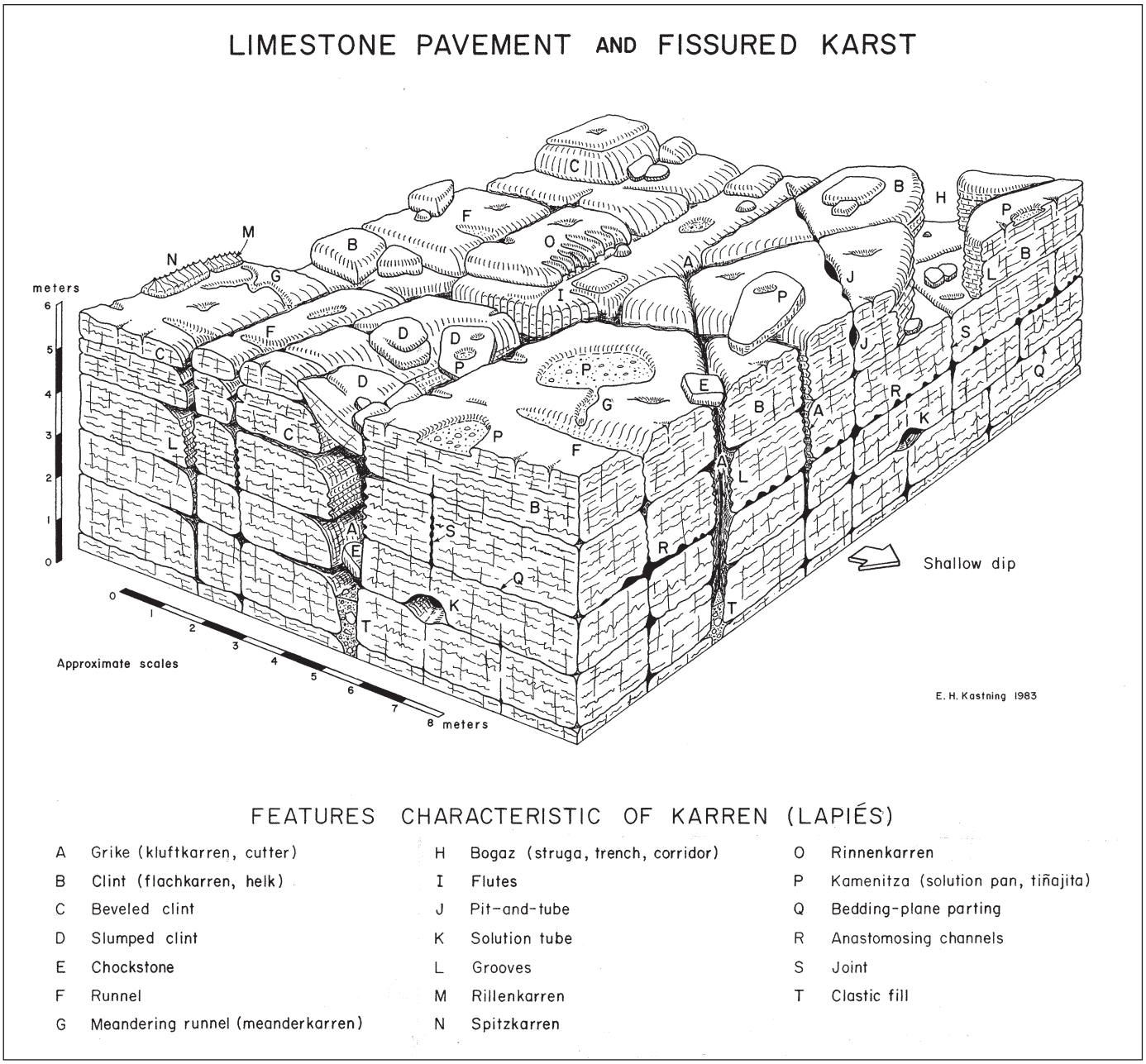


Figure 98: Generalized block-diagram of the McLarrin Fissure Karst showing solutionally enlarged joints and bedding-plane partings, pavement on limestone, geologic structure, dislodged clints, and flow of vadose seepage.

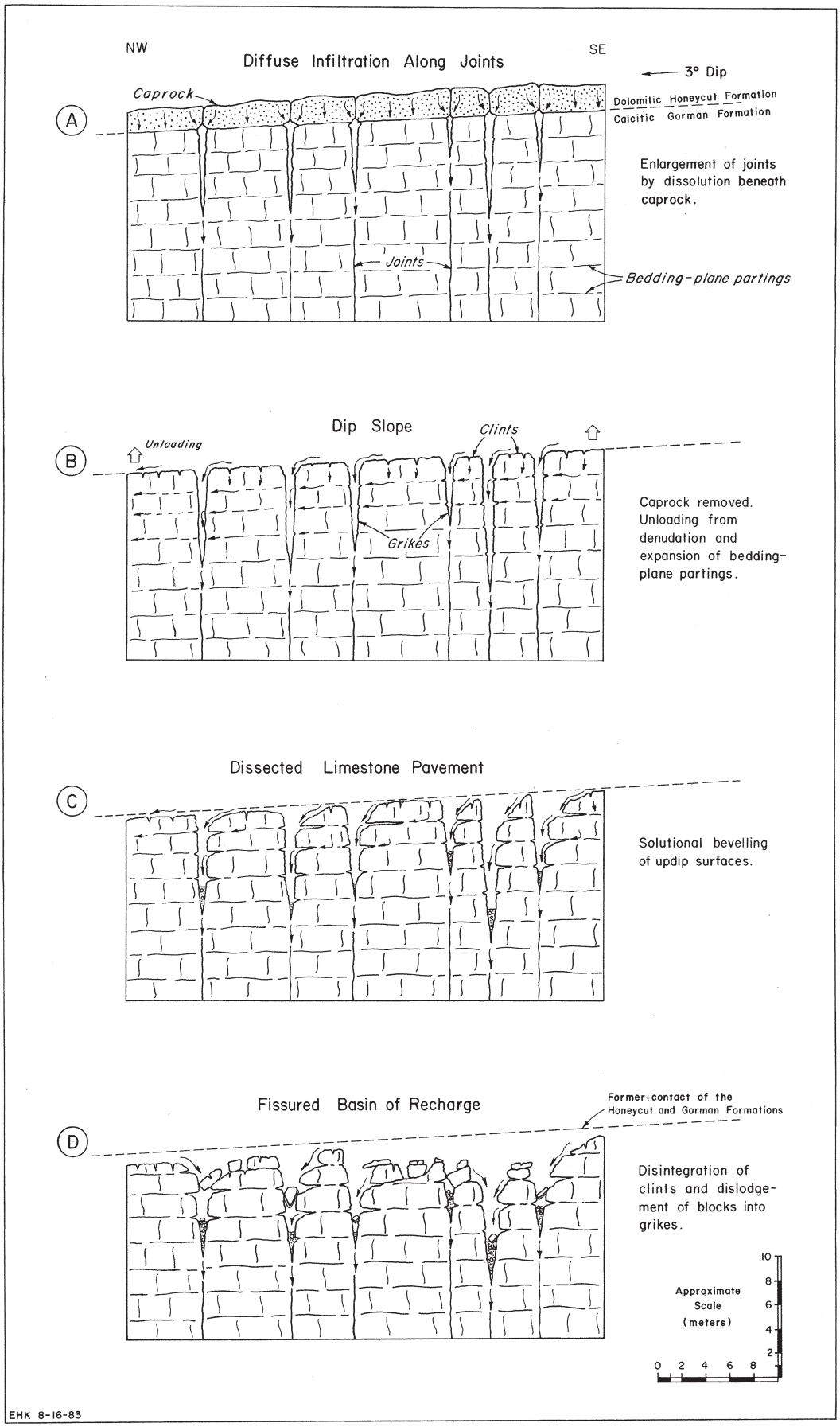


Figure 99: Origin of fissure-like caves. (A) Solutional widening of fractures in limestone (Gorman Formation) between cover-beds of dolostone (Honeycut Formation). (B) Removal of overburden and expansion of bedding-plane partings updip from vertical fractures; and bevelling of horizontal surfaces into fissures. (D) Undercutting and disintegration of clints; dislodgement of blocks over and into fissures.

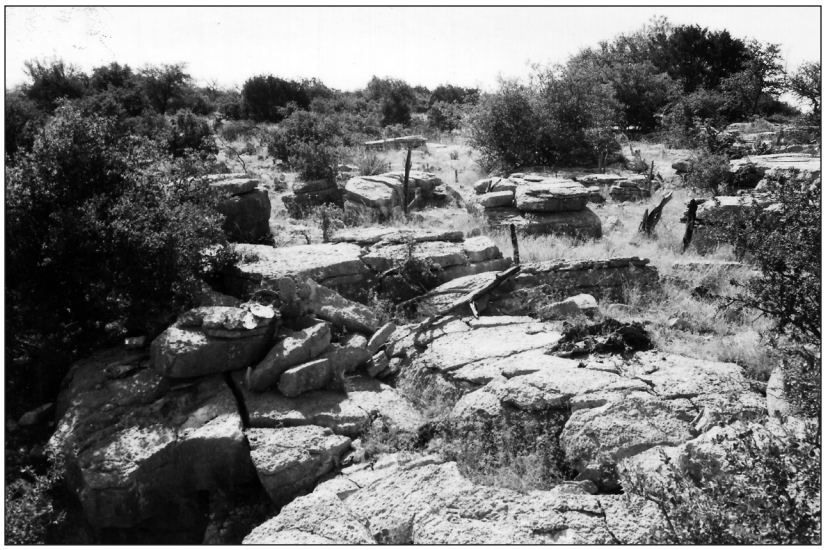


Figure 100: Advanced stage of the McLarrin Fissure Karst. Dagger Cave (Figure 85L) is at lower left. Clints have undergone disintegration, displacement, and slumping. See also Figure 99D.

Origin of Caves with Single Rooms

Short, horizontal caves, such as Copperhead, Debris, and Wedge Caves and Gorman Sink, are small, dry, and consist of single rooms. Because they occur in beds of limestone near the surface, they are viewed as isolated pockets of solution similar to many small caves elsewhere on the Edwards Plateau. They are older than most other caves of the Bend area, having formed when baselevel and the piezometric surface were considerably higher.

Origin of Pit-Caves

Deep pits, including El Sotanito Grande, El Sotano Lodoso, and Yates Goat Cave, developed along vertically extensive fractures that communicated water from points of infiltration on the surface directly to lower, cavernous zones. Presumably horizontal conduits exist at or below their lowest points, but these are now blocked by fill. Fractures were widened by sheet-flow along walls, dissolving the bedrock in a manner similar to the development of vertical shafts. These pits may have been capped by beds of dolostone during their initial development.

Origin of Fissure-like Caves

Caves of the Gorman Fissure System and McLarrin Fissure Karst (Figure 98) represent large-scale grikes or Kluftkarren (solutionally widened fissures in a pavement of limestone). They may have been covered at the onset of development by resistant beds of dolostone, especially in the case of the McLarrin fissures (Figure 99A). Seepage from the surface penetrated the beds of dolostone along vertical fractures, and, upon reaching the limestone beneath, began to enlarge the fractures through solutional widening. As the overburden and caprock were removed, runoff from the limestone-pavement was able to enter the fissures directly. Removal of the overlying strata and consequent unloading allowed beds near the surface to part along bedding planes (Figures 83, 84, 99B, and 100). Some seepage moved down the dip along bedding-plane partings to intersections with grikes. Clints (limestone blocks between grikes) became solutionally etched on their upper surfaces and bevelled downward into the grikes at their downdip margins (Figures 84 and 99C). Lower surfaces of bedding-plane partings were similarly modified, leading to the development of small niches and shelves along the up-dip walls of fissures. Clints near the surface were solutionally modified by widening of vertical joints and undercutting by lateral flow and dissolution (Figure 100). Some loosened blocks have become dislodged and many have slid over fissures or fallen into them (Figures 84, 86, 87, and 99D). This process is still occurring, as shown by clints with minor karstic forms on their upward-facing surfaces, now tilted by slumping into fissures (Figure 101). Smaller blocks, other weathered debris, and vegetal matter have littered the floors of all fissures, blocking further exploration. Based on these observations, the fissure-like caves appear to be relatively young karstic features, dating to the late Pleistocene at the oldest.

Origin of Caves in Travertine

Caves in travertine described in this chapter and others in the Llano region are not true karstic features and may best be described as pseudokarstic. They have largely formed by accretion rather than by dissolution. Travertine is produced by precipitation of calcite as carbon dioxide is lost from a solution saturated with respect to calcite (Pitty, 1971). Of the factors responsible for depletion of carbon dioxide, those which are pertinent here include (1) rise in temperature, (2) agitation, and (3) use of carbon dioxide by aqueous plants such as algae. It is likely that all three contributed to deposition of travertine in the Bend area. Water issuing from Gorman Spring flows along Gorman Creek for about 900 m before reaching Gorman Falls. In this reach water is warmed and agitated as it cascades over small ledges. Numerous dams of travertine 5 to 30 cm high occur along the creek. Algae and other small, green plants abound along the creek as well. The greatest agitation occurs at Gorman Falls where deposits of travertine are most massive. Deposits vary from 2 to 15 m in thickness and rest unconformably on the Gorman Formation (Figures 90 and 102). As water cascades over the lip of the terrace of travertine, it loses carbon dioxide and deposits calcite outwardly from the lip, forming an overhang. At the same time calcite is deposited at the base of the falls by water splashing upon impact and evaporating (Figure 103A). The overhang continues to grow out and down as a canopy of stalactitic deposition (Figure 103B), and stalagmitic deposition at the base may eventually enclose a niche, forming a small cave (e.g. Mini Cave). Deposits from above and below eventually coalesce, forming a natural tunnel which may later be partially closed at the ends and gradually filled in with speleothems deposited by dripping water (Figures 82 and 103C).

Deposition of travertine has been occurring at Gorman Falls for some time. The spillover point of Gorman Creek has gradually shifted to the northwest. Deposits within 150 m southeast of the falls have been abandoned by Gorman Creek and are desiccating. Some overhanging deposits have collapsed, perhaps in response to erosional undercutting by high stages of the Colorado River. Large blocks of travertine have fallen into the river, deflecting it slightly to the northeast.

Chronology of Speleogenesis

The following chronological sequence is proposed for the origin of caves in the Bend area:

1. The carbonate units of the Ellenburger in the Llano region were karsted from the late Ordovician through the Mississippian, and preserved structures of collapse contain sedimentary strata of these ages. At least three collapsed areas of the Ordovician have been mapped in the Bend area; however, younger infillings have not been discovered.

2. Major faulting accompanied uplift of the Llano region during the Pennsylvanian, partitioning the Bend area into several fault-blocks. Fairly uniform sets of

Figure 101: Recently slumped fragment of clint over Snail Fissure, McLarrin Fissure Karst (Figure 85-O). Note Kamenitza on formerly up-facing surface. Pencil is approximately 17 cm long.



fractures developed concurrently with faulting, and fault-blocks became tilted at various angles.

3. Uplift of the Edwards Plateau during the Miocene initiated erosion of uplands and incision of the Colorado River. This was followed by development of an erosional surface over the area prior to the mid-Pleistocene and then rejuvenation. This led to incision of the Colorado and its tributaries through the carbonate beds of the Ellenburger Group and development of a local baselevel toward which groundwater could migrate.

4. Drainage developed above resistant beds of dolostone in the direction of dip. This established dip-slopes on the surface, especially in the northwestern areas. In the subsurface, conduits carrying groundwater through the limestone became perched on beds of dolostone.

5. Enlargement of caves ensued in dense, highly-pure limestone of neglible, primary porosity. The limestone was dissolved along vertical fractures, with the greatest development of master conduits down the dip toward the Colorado River in the southeastern part of the Bend area.

6. Gorman Cave, the largest cave in the area, formed at a time when the bed of the Colorado River was 12 to 15 m above its present elevation. Perched on a bed of dolostone, the cave became graded to that baselevel. Most bluff-caves along the Colorado River developed at this time.

7. Constrictions in Gorman Cave at faults and brecciated zones promoted impoundment of water upstream of those points and consequent enlargement by floodwater. Velocities of flow were great enough through the constrictions to induce highly turbulent flow which then sculpted walls, floors, and ceilings of passages.

8. The stream through Gorman Cave eventually breached the bed of dolostone beneath the main (upper) level in response to further entrenchment of the Colorado River. Lower-level passages carrying the stream became perched on another unit

Figure 102: Deposit of travertine resting uncomfortably on Gorman Formation near Gorman Falls. Hammer (33 cm long) is at contact.



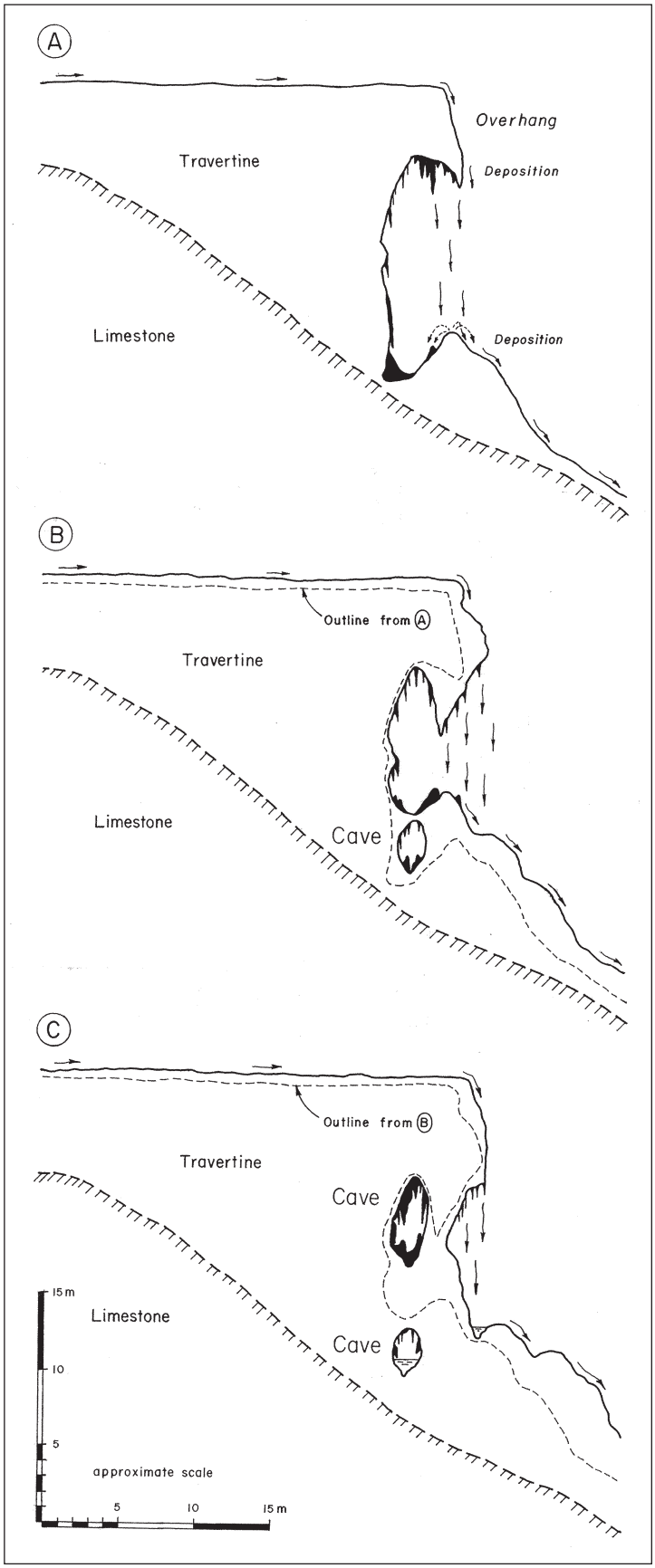


Figure 103: Origin of caves in travertine. (A) Development of overhang from precipitation of calcite and growth of mound of stalagmitic deposits at base of falls. (B) Continued growth of these deposits, forming caves beneath canopy and in niche at base of mound. (C) Eventual enclosing of cave and infilling with speleothems.

of dolostone.

9. Gorman Cave experienced periodic flooding during the late Pleistocene and Holocene, especially during cool and wet pluvial climates when discharge of the river was presumably much greater than it is today. Floods enlarged passages slightly and deposited medium- to coarse-grained terrigenous sediment along the floors of passages.

10. Erosional removal of the Honeycut Formation from hills in the northwestern part of the area exposed solutionally enlarged crevices of the McLarrin Fissure Karst. Runoff flowing down the incipient dip-slope was captured by fissures and shallow bedding-plane partings. This contributed to enlargement of fissures to their present widths and depths. The Gorman Fissure System may have formed in a similar manner.

SUMMARY

The Bend area is underlain by a sequence of alternating limestone and dolostone that has been deeply dissected by the Colorado River and its tributaries. Although the carbonate beds are high in density of grains and low in primary porosity, they are high in content of calcite, and speleogenesis has been strongly influenced by fractures. Infiltrated water from the uplands was guided vertically along fractures to depths at which resistant beds of dolostone were encountered. Mature conduits evolved in beds of limestone where perched groundwater flowed down the dip along open fractures toward baselevel at the Colorado River. The caves evolved conformably with the strata and were later modified by flooding behind constrictions at faults and zones of collapse and by backflooding during rises in the Colorado River.

Fissure-like caves consist of large-scale grikes developed where closely-spaced and open fractures have been solutionally enlarged. Denudation, removal of overlying dolomitic beds of the Honeycut Formation, and unloading, allowed expansion of fractures and bedding-plane partings, permitting down-dip flow along the latter and enlargement of low, horizontal passages. Bluff-caves along the river represent either isolated remnants of former conduits long since abandoned or short spurs excavated by backflooding of the river when baselevel was higher than it is today. Caves in travertine at Gorman Falls evolved by accretion of calcite brought on by agitation and the loss of carbon dioxide as water cascaded to the Colorado River.

9
BURNET AREA

DESCRIPTION OF AREA

The Burnet area of the eastern Llano Basin lies generally east of the Colorado River in west-central Burnet County but extends slightly west of the river into east-central Llano County (Plate 1). As in the Bend area 41 km to the north (Chapter 8), karst has developed on carbonate rocks of the Ellenburger Group (Ordovician). The Burnet area includes Longhorn Cavern, the seventh longest and one of the best known caves of Texas. The Burnet area is bounded on the south and north along lines of 30°37'30" and 30°45'00" north latitude respectively, and on the east and west along lines of 98°17'30" and 98°25'00" west longitude respectively (Figure 104). The area consists of the western two-thirds of the Longhorn Cavern and eastern third of the Kingsland 7.5-minute U.S.G.S. quadrangles. Drainage is to the west and south, to the Colorado River and its tributaries, including Peters and Backbone Creeks (Figure 104).

Published geologic maps of the Burnet area range in scale from 1:31,680 to 1:250,000 (Paige, 1912a; Barnes, 1956a, 1981a; Semken, 1960b; Matthews, 1963; Boyer and McQueen, 1964). Maps that show nearby outcrops of the Ellenburger and their relationships to other units and regional structure include those of Cloud and Barnes (1946a, figures 7 and 8, plates 10 and 11), Goldich and Parmelee (1947), Barrow (1948), McGehee (1963), McQueen (1963), Boyer and McQueen (1964), Iranpanah (1964), Groshong (1967), Reid (1968), Barnes and others (1963, figures 3, 11, and 12; 1972, figures 3, 17, and 18) and Barnes (1976b).

The karst of the Burnet area is contained within the wedged-shaped Backbone Ridge Graben, 0.6 to 11.3 km wide and approximately 19 km long, trending northeast from Kingsland to Burnet (Figure 104). The graben is bounded on the northwest by the Roaring Spring Fault Zone and on the southeast by the Bald Knob Fault Zone. Units of the Moore Hollow and Ellenburger Groups are exposed on the graben. The Riley and Wilberns Formations crop out at the southwestern end of the graben and along the southeastern flank of its northeastern part. The Tanyard, Gorman, and Honeycut Formations are exposed in the remainder of the graben. The immediate vicinity of Longhorn Cavern (Figure 105), includes nearly the entire sequence of the Ellenburger (Plate 4 and Figure 8).

Geologic sections of the Ellenburger and localities for building-stone within the Burnet area are described in Barnes (1936), Cloud and Barnes (1946a, p. 287-308), Barnes, Dawson, and Parkinson (1947, p. 141, 142; 1963, p. 113-117; Barnes and others, 1972, p. 54, 56-59), and Dietrich and Lonsdale (1958, p. 25, table 5). In general, the lithology and thickness of the Ellenburger units in this area are comparable to those in the Bend area of Chapter 8 (Figures 8 and 88). The following description of Ellenburger units on Backbone Ridge, approximately 3.7 to 5.8 km southwest of Longhorn Cavern, is summarized from Cloud and Barnes (1946a, p. 291-292).

The Threadgill Member of the Tanyard Formation is a coarse-grained, white to light gray dolostone which, in places, grades to limestone. A small amount of chert occurs in the upper part. The thickness of the Threadgill is 28 m at this location. The overlying Staendebach Member of the Tanyard Formation is predominantly a dolostone, medium- to fine-grained in the lower part and medium- to coarse-grained in the upper part; however, locally the Staendebach grades to limestone. Porcelaneous to semiporcelaneous dolomoldic chert is abundant in the unit. Thickness of the Staendebach is 146 m at this site.

The Gorman Formation of the Burnet area consists of a lower dolomitic facies, 67 m thick, and an upper calcitic facies, 80 m thick. The dolomitic facies is mostly microgranular dolostone with some fine- to medium-grained dolostone and some limestone in places. The calcitic facies consists of an upper and lower part composed predominantly of aphanitic, thick- to thin-bedded, light gray limestone and a middle part which is typically microgranular dolostone. This tripartite division is similar to that observed in the Bend area (Figures 8 and 88). The upper 16 m of the Gorman Formation consist of very massive, pure limestone. The *Archaeoscyphia* zone lies approximately 13 m above the contact between the dolomitic and calcitic facies (Figures 8 and 88).

The overlying Honeycut Formation in the Burnet area consists of alternating thick- to thin-bedded limestone and dolostone, with an overall preponderance of microgranular dolostone. It is generally more dolomitic at this locality than in the Bend area (Figure 88) and contains considerable chert and fossils. The thickness of the Honeycut in the Burnet area is indeterminable due to erosional truncation. Additional remarks on the Ellenburger Group, including references to the literature, are provided in Chapters 2 and 8 and in Appendix B.

The graben composed of rocks of the Moore Hollow and Ellenburger Groups is flanked on the northwest by exposures of Town Mountain Granite and Valley Spring Gneiss (both Precambrian), and to the southeast by the Town Mountain Granite and units of the Moore Hollow Group (Figure 104). Quaternary alluvium occurs along the Colorado River and major creeks of the area, and small patches of travertine occur along the escarpment formed by the Roaring Springs Fault Zone (e.g. Cloud and Barnes, 1946a, figure 8, p. 289).

Faulting and development of the Backbone Ridge Graben occurred during epeirogeny in the Pennsylvanian (Powers, 1928; Sellards, 1934, p. 85; Cloud and Barnes, 1946a, p. 120-121; Eardley, 1951, p. 228; Cheney and Goss, 1952; also see Chapter 2 for summary of Paleozoic tectonism). Under tensional stress the overlying Cambrian and Ordovician sedimentary rocks dropped between Precambrian igneous and metamorphic rocks (Figure 106A,B). During this event differential displacement occurred along the Roaring Spring Fault Zone. Total throw was a maximum (610 m) about 1.6 km northeast of Roaring Spring (Boil Spring on recent maps) and declined to the northeast and southwest (Cloud and Barnes, 1946a, p. 290 1946a, p. 290). As a result, Paleozoic strata within the graben have been folded into a gentle syncline plunging approximately 7° to the northwest, with dips along its flanks ranging between 2° and 18° (Figures 104 and 105).

Fractures in the Ellenburger of the Backbone Ridge Graben occur in distinct sets that correspond to faults and synclinal folding. This agreement is substantiated by maps of joints and aerial photolineaments by McQueen (1963), Boyer and McQueen (1964), and Reid (1968, 1969). Fractures are spaced a few meters apart on the average and in the vicinity of Longhorn Caverns appear to be more numerous where the calcitic member of the Gorman Formation overlies caverns and solutional openings (Barnes, Dawson, and Parkinson, 1947, p. 141-142). Away from cavernous zones this member is more massive with a lower density of fractures.

Today Backbone Ridge, bordered by escarpments along the Roaring Spring and Bald Knob Fault Zones, is an upland surface lying generally 75 to 150 m above the surrounding lowland. The Town Mountain granite weathers readily and erodes to a generally flat surface (Barnes, 1956a, p. 12), whereas beds of the Ellenburger are relatively resistant. As a result, the fault-block containing them is left standing high over the surrounding area (Figure 106C,D,E). The implications of differential erosion and inversion of topography on the origin of Longhorn Cavern are discussed later in this chapter.

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

Llano and Burnet Counties have about 65 known caves, of which only 4 exceed 300 m in length (one of these is Enchanted Rock Cave, a pseudokarstic feature discussed in Chapter 13). Longhorn Cavern (Figure 107) on Backbone Ridge is the longest mapped cave in this region and one of the best known in Texas.

Longhorn Cavern, previously called Hoover's Valley Cave or Sherrard Cave, has been known for some time and is rich in history. Archaeological evidence indicates that the cave was visited and used by Indians and that it once served as a refuge for outlaws (Craun, 1948; Matthews, 1963; Maxwell, 1970). During the Civil War, Confederate workers mined, leached, and stored nitrates derived from deposits of bat-guano in the cave for use in manufacturing gunpowder (Phillips, 1901; Meador, 1964). Later the cave was operated as a nightclub (Matthews, 1963). As indicated on the first map of the cave (Law, 1933, figure

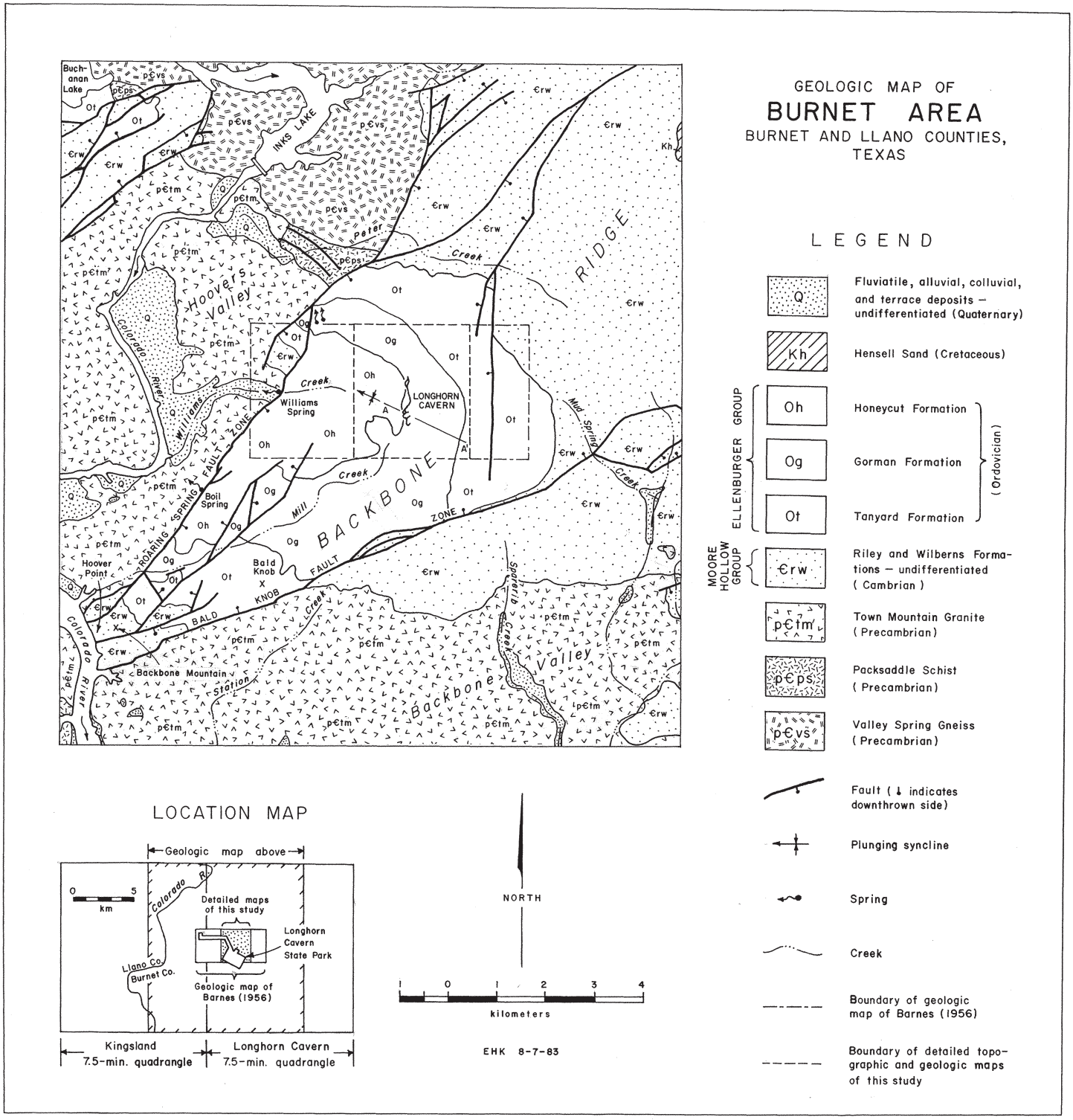


Figure 104: Geologic map of the Burnet area, Burnet and Llano Counties. Detailed maps of area inside dashed line are presented in Figures 105 and 108. Geology modified from Paige (1912a) and Barnes (1956a, 1981a).

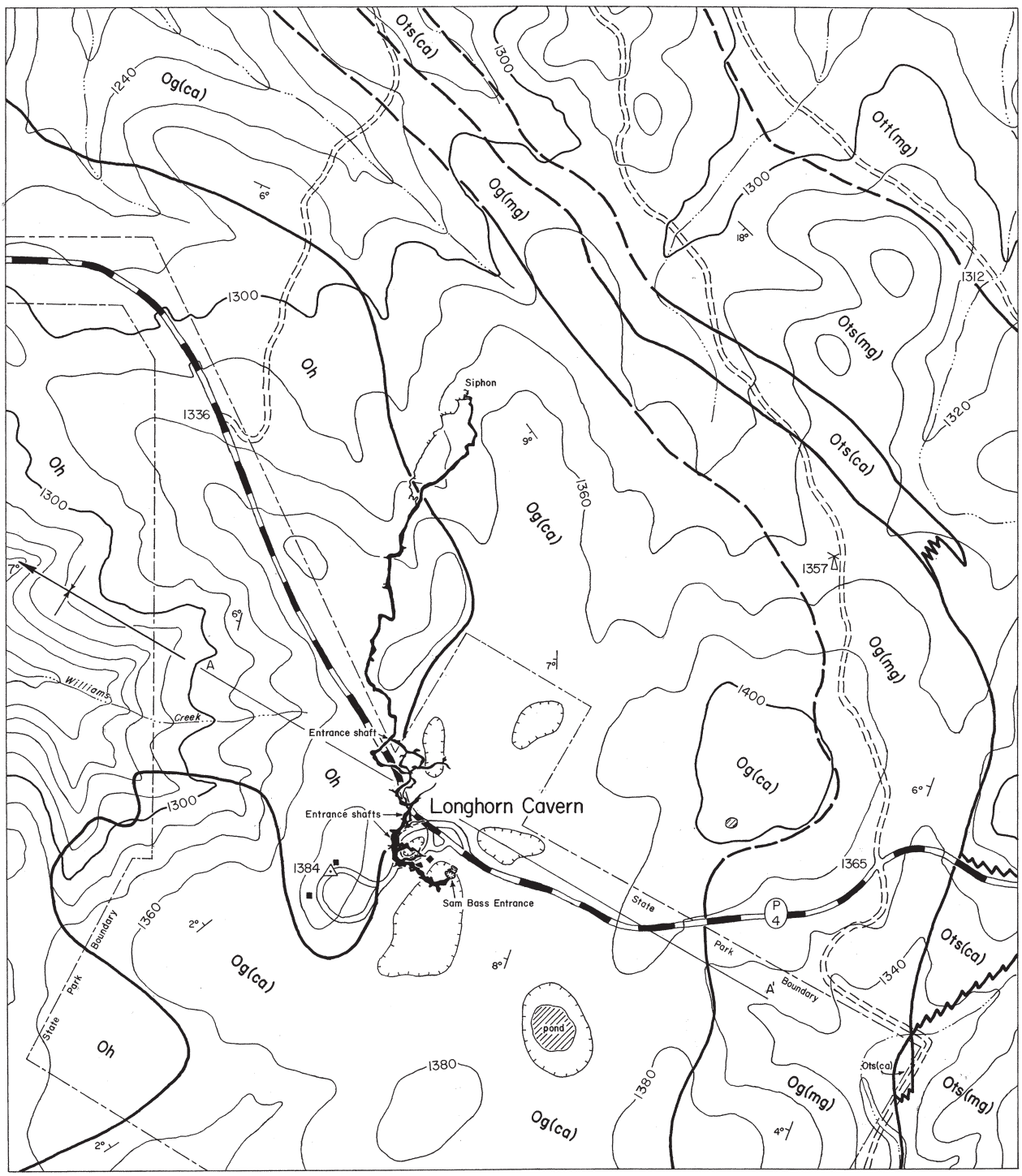


Figure 105: Topographic and geologic map of the vicinity of Longhorn Cavern showing structural features and the cave. Geologic section A-A' is shown in Figure 109 see Figure 104 for location of area). Geology modified from map of Barnes (1956a, p. 18).

1, p. 35), tours initially used the Main (now sealed) and the San Bass Entrances and visited the Cathedral Room and the Hall of Marble (Figure 107). However, the route between these two chambers was through the lower-level Lake Room Tunnel, because the overlying passages, from the Chandelier Room to the Hall of Wonders, were partially occluded by sediment. Considerable debris, including earth, guano, and rock, was excavated form the cave in the late 1930's by the Civilian Conservation Corps in order to open these and additional passages for tours (Craun, 1948; Matthews, 1963; Semken, 1960b, 1961; Shore, 1964). In all, approximately 15,000 m^3 of material was removed from the cave, much of it through an artificial shaft between the surface and the junction of the Hall of Marble and Lumbago Alley (Figure 107).

Law (1933) investigated the geology of the cave, surveyed the readily accessible passages in plan and profile, and discussed the origin of the cave in context with the views on speleogenesis prevailing at that time. His paper caught the attention of William Morris Davis, who briefly commented on these observations (Davis, 1933). As part of their study of the Ellenburger Group Cloud and Barnes (1946a, P. 51, 132, 292) noted that the cave had developed in the uppermost, massive beds of the Gorman Formation. Later, Craun (1948) further described the cave and reprinted Law's (1933) paper, and Barnes (1956a) mapped the geology of the vicinity.

Modern investigations began with projects of the Texas Speleological Association (T. S. A.) held at the cave in 1957 and 1958 (anonymous, 1956; Jackson, 1957, 1958, 1959). Matthews (1963) authored a popular geologic and historic guidebook to the cave which included a crude map, and Reddell (1964b) and Maxwell (1970) provided brief descriptions. A third T.S.A. project in 1971 produced an accurate map (Figure 107) showing nearly all known passages, but this was not published until seven years later (Fieseler and others, 1978, p. 85).

Longhorn Cavern is an important locality for vertebrate fossils. Semken (1960a,b, 1961) collected and inventoried Late Pleistocene fossils along the main trail through the cave. He identified a profuse population of rodents and found several species of higher vertebrates no longer endemic to central Texas. He also mapped four stratigraphic units in the sediments of the cave. Kurten (1963) studied fossils of bears and Patton (1963a) reported on remains of southern bog lemmings in the Pleistocene sediments of the cave. Frank (1961a,b, 1964, 1965), Lundelius (1967), Lundelius and Slaughter (1971b) included many of these findings in summaries and syntheses of the paleontology of vertebrates from the caves of Texas.

Hydrogeologic investigations of the carbonate rocks of the Burnet area have been few. Brief comments on the Ellenburger aquifer were made by Plummer (1946) and Mount and others (1967). Mount (1962) investigated conditions for groundwater in the vicinity of Burnet, but his study did not extend southwest to the Longhorn Cavern.

DEVELOPMENT OF CAVES

Description of Longhorn Cavern

Longhorn Cavern is 6.1 km southeast of the Roy Inks Dam on the Colorado River and 4.0km east along Park Road 4 from its junction with Route 2342 (Figure 104). The Sam Bass Entrance, in a doline 60 m south of Park Road 4 (Figure 105), is marked on the Longhorn Cavern 7.5-minute U.S.G.S. quadrangle. The cave has

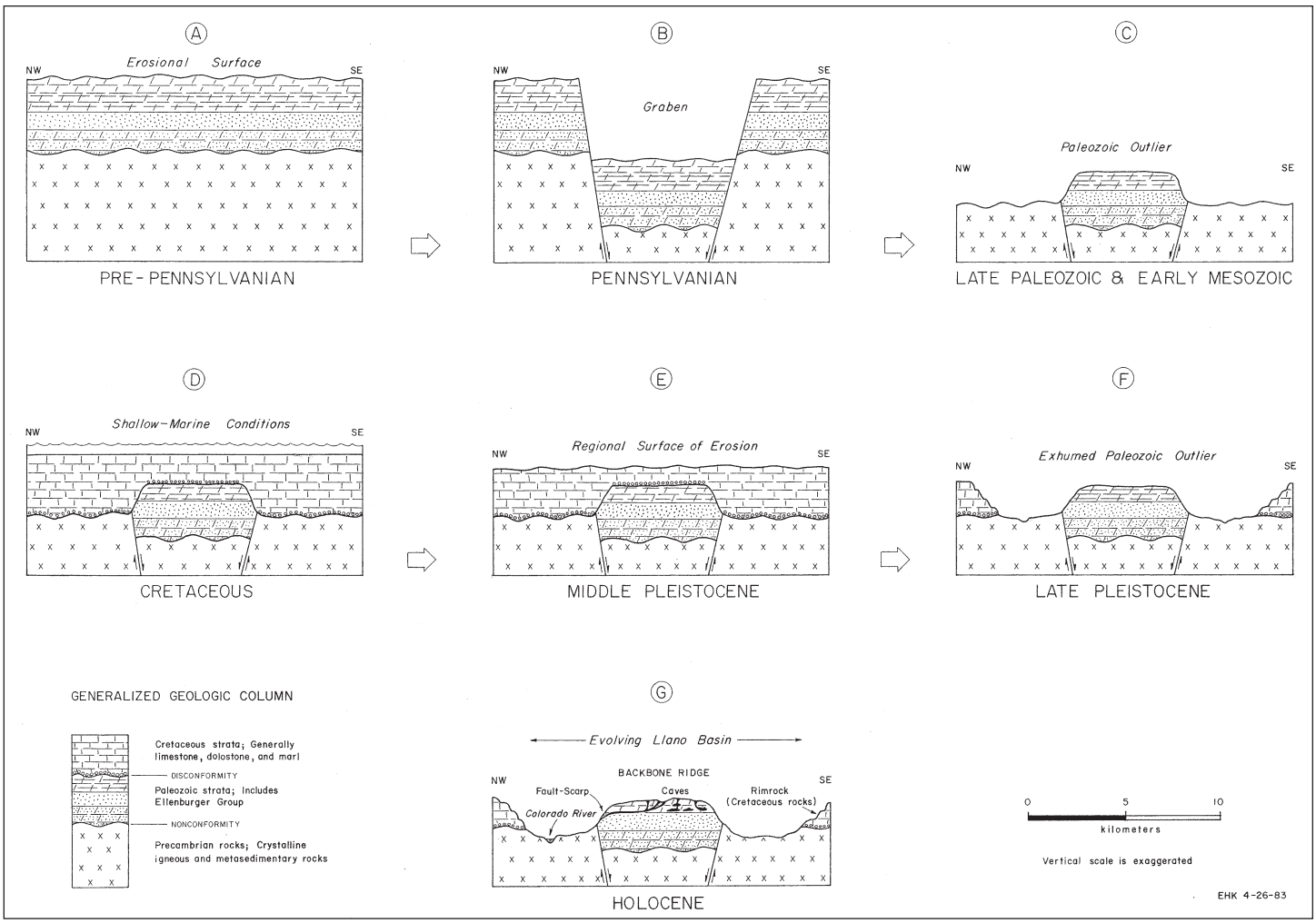


Figure 106: Schematic diagram explaining origin of Backbone Ridge and Longhorn Cavern: (A) Original stratigraphic sequence of Paleozoic strata overlying Precambrian crystalline rocks, (B) Block-faulting during the Pennsylvanian, resulting in a graben where Paleozoic rocks are juxtaposed against Precambrian rocks, (C) Differential erosion during the late Paleozoic and early Mesozoic resulting in inversion of the topography with respect to structural displacement, (D) Deposition of shallow-marine, carbonate rocks during the Cretaceous, (E) Denudation during the late Tertiary to mid-Pleistocene, forming an erosional surface over the area, (F) Removal of Cretaceous sediments during the remainder of the Pleistocene and exhumation of the pre-Cretaceous outlier at Backbone Ridge, (G) Developemnt of caves in response to a reduction in baselevel adjacent to the outlier. Based partly on Stenzel (1937, p. 3-4)

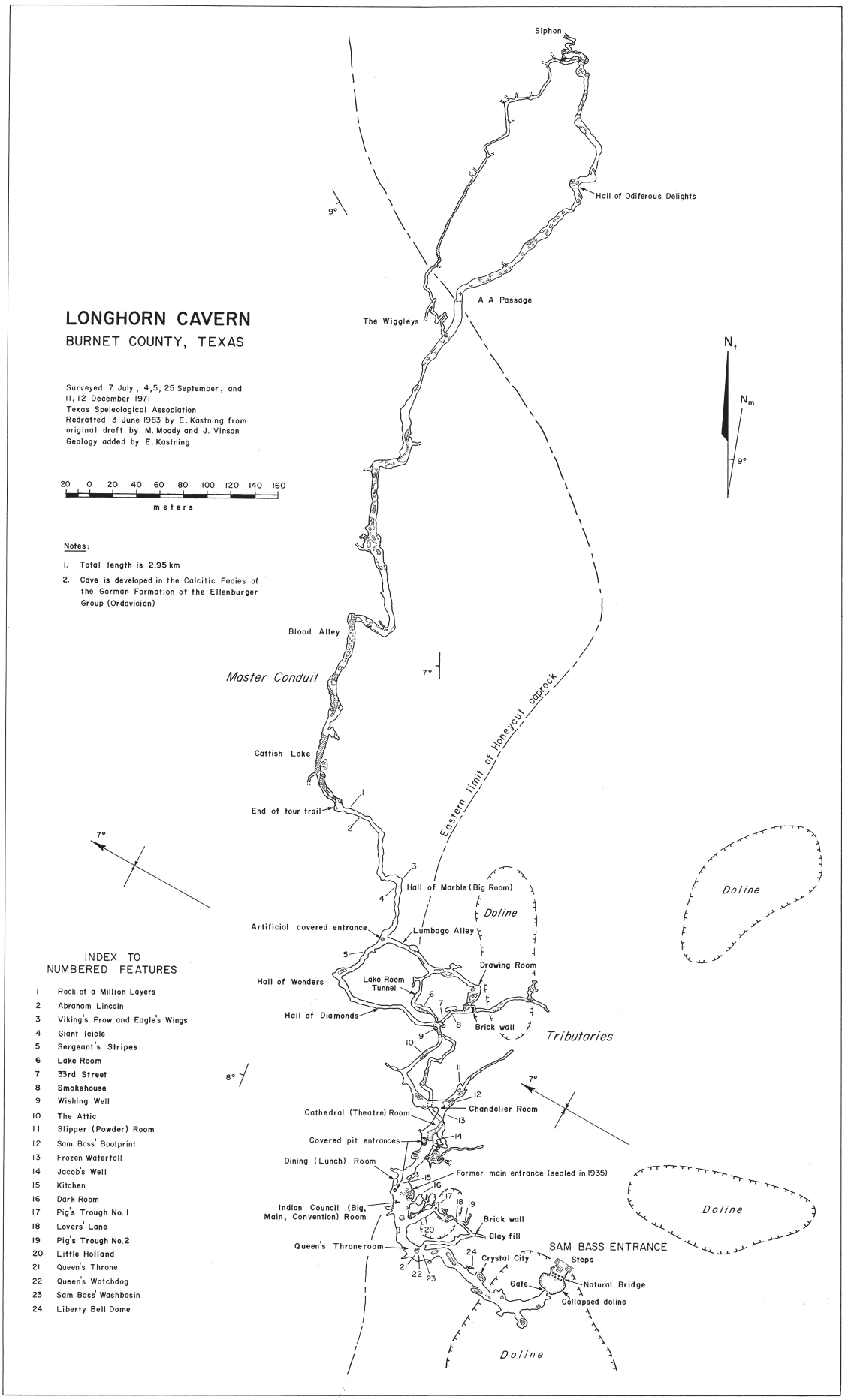


Figure 107: Map of Longhorn Cavern. Modified from Fieseler and others (1978, p. 85). Geographic names added from Law (1933) and Matthews (1963)

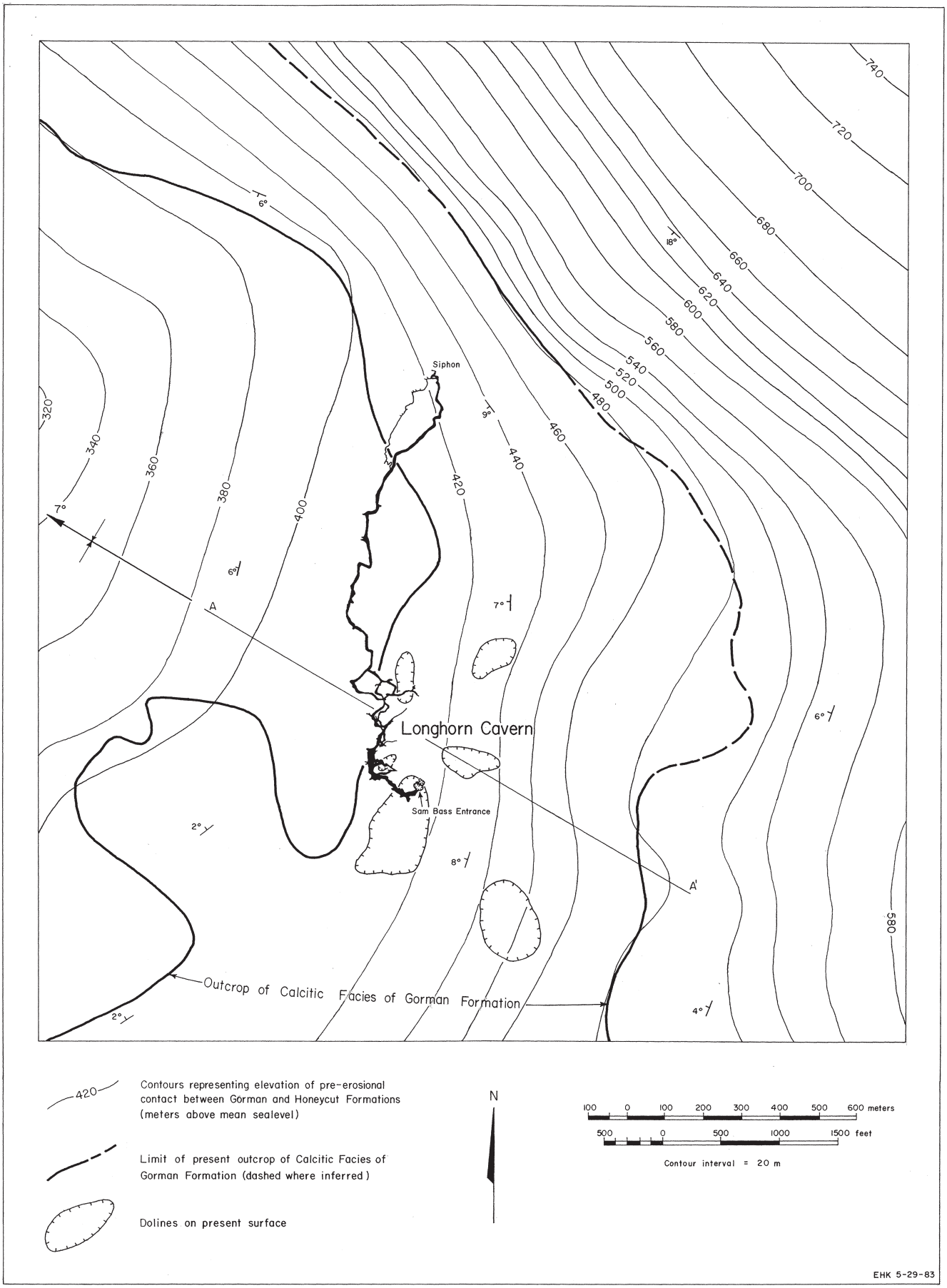


Figure 108: Structural contour-map of the pre-erosional contact between the Gorman and Honeycut Formations in the vicinity of Longhorn Cavern. Geologic section A-A' is shown in Figure 109. Area is identical to that of Figure 105 (see Figure 104 for location). Map was constructed from U.S.G.S. Longhorn Cavern 7.5-minute quadrangle (1967), geologic map of Barnes (1956a, p. 18), and stratigraphic informaion in Cloud and Barnes (1946a, p. 287-308).

a mapped length of 2.95 km, an end-to-end extent of 1100 m, and internal relief of approximately 4m (Figure 107).

There are four natural and one artificial entrances to the cave. The Sam Bass Entrance, presently serving as the entrance and exit for tours, is a collapsed doline 10 m deep and approximately 23 m in diameter. A remaining span of rock forms a natural bridge over the doline. The Main Entrance, located 140 m to the northwest, was once open to another doline but was sealed in 1935. Two natural entrances (shafts) approximately 18 m deep occur 192 m and 240 m respectively along the main passage from the Sam Bass Entrance. An 18 m-deep artificial shaft intersects the main passage 535 m from the entrance, at the junction of Lumbago Alley and the Hall of Marble. It was excavated to facilitate removal of fill during commercialization.

Longhorn cavern consists of three major sections (Figure 107). The southern third of the cave is a complex of passages consisting of (1) a master conduit extending from the Sam Bass Entrance to the Hall of Marble, (2) a lower-level conduit (Lake Room Tunnel) from the Cathedral Room to Lumbago Alley, and (3) numerous short passages branching eastwardly from the main conduit. This gives the southern end of the cave a dendritic and maze-like pattern. The middle third of the cave consists of a single passage with a few short side-passages and is an extension of the master conduit from the Hall of Marble to the AA Passage. The northern third of the cave consists of the final segment of the main conduit and a smaller, branching and parallel passage to the west. Both passages connect near the siphon at the north end of the cave and also near the Wiggleys.

Passages are generally meandering, but most consist of linear segments, suggesting structural control. Along its 1.56 km length the main conduit averages between 2 and 10 m in width and from 1 to 6 m in height. These dimensions reach 15 and 10 m respectively in large chambers (e.g. Indian Council Room, Chandelier Room). In many places domes in ceilings locally increase the height of passages by as much as 3 or 4 m.

The southern third of Longhorn Cavern lies generally within the upper, massive beds of limestone of the calcitic facies of the Gorman Formation (Cloud and Barnes, 1946a, p. 51, 133, 292). However, much of the Lake Room Tunnel and passages of the northern two-thirds of the cave are developed along 1 to 3 m-thick beds of limestone of an otherwise thick, dolomitic zone within the calcitic facies (Figure 88). All natural entrances to the cave lie within the outcrop of the calcitic facies, at or just east of the outcrop of the Honeycut Formation (Figure 105). The artificially excavated shaft pierces the lowest Honeycut beds just west of the Gorman-Honeycut contact.

Longhorn Cavern does not have any perennial flow of water. During wet seasons water enters along tributary-passages on the east flank of the cave and lower levels become muddy and are often flooded to the ceilings, forming siphons (Reddell, 1964b, p. 44; Fieseler and others, 1978, p. 87). Water also enters the cave as vertical seepage along joints. Water in the lower passages of the cave is discharged through the siphon at the present northern limit of exploration (Figure 107). Numerous shallow pools (e.g. Catfish Lake) remain throughout the year.

The cave is floored with sediment ranging in size from silt to pebbles. Deposits of breakdown are few and are principally associated with areas near entrances. Much of the fill in the southern third of the cave was removed during commercial development or covered by trails, and consequently *in situ* deposits are visible only in a few places in this section.

Speleothems are limited to deposits of flowstone and large crystals of calcite. Flowstone and dripstone, occurring throughout the cave, are no longer being actively deposited, and modest displays of draperies, stalactites, and stalagmites are visible in many alcoves and pockets in ceilings (see photographs in Matthews, 1963). Massive deposits of flowstone (e.g. the Queens Throne) are particularly abundant in the southern third of the cave on both levels and in side passages as well as in the main passage. One such deposit, the Rock of a Million Layers at the end of the trail used for tours (Matthews, 1963, figure 19, p. 23), is a fine example of redissolution. Thick deposits of crystals of calcite line the walls of several passages, including the Crystal Rooms and Hall of Diamonds. Individual crystals are typically 5 cm in diameter or larger.

Longhorn Cavern lies beneath a bench or tableland developed on the upland of Backbone Ridge. The immediate vicinity of the cave is of low (35 m or less) relief. However, the tableland has been deeply dissected 3.0 km to the northwest and 2.4

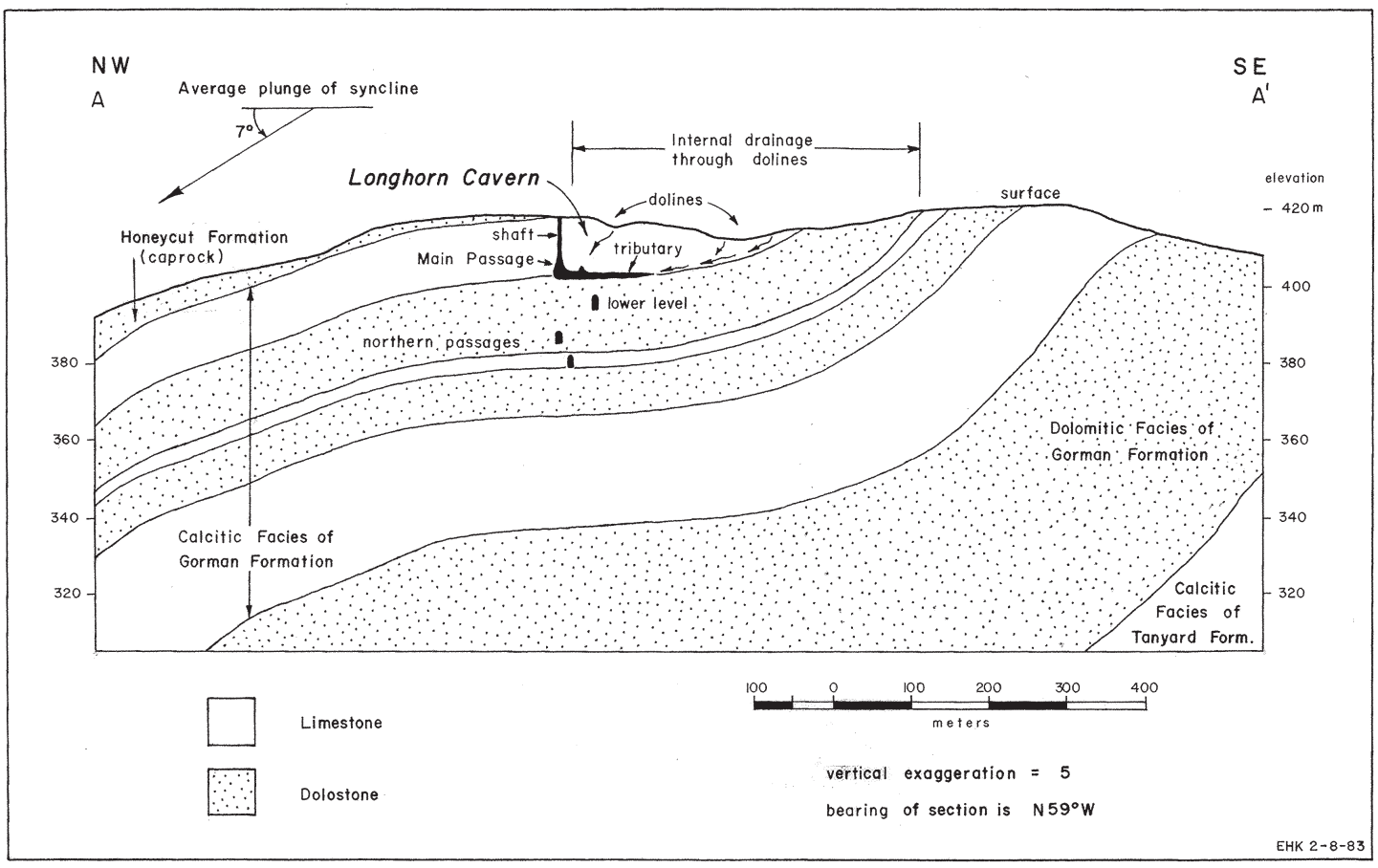


Figure 109: Geologic cross-section A-A' along plunge of synoclinal axis at Longhorn Cavern showing relationship of the cave to strata and dolines. Location of section is shown in Figures 105 and 108. Constructed from same sources listed in caption to Figure 108.

km to the southeast along the fault scarps that delimit the Backbone Ridge Graben (Figure 104). Relief between the entrances to the cave and the lowlands below the scarps is approximately 100 to 150 m. The cave lies just northwest of the drainage divide along Backbone Ridge from which surficial runoff flows toward the scarps to the northwest, south, and southeast.

Speleogenesis

Five principal relationships exist between Longhorn Cavern and the local topography and geology: (1) passages lie on four levels and are generally concordant with strata of the Gorman Formation, (2) the cave is centrally positioned beneath a topographic tableland bounded by scarps at faults (Figures 104 and 105), (3) the area around the entrance of the cave is associated with an isolated cluster of dolines developed within the outcrop of the Gorman Formation (Figure 105), (4) the main conduits lie along the nose of a synclinal fold and trend parallel or subparallel to the strike of the bedrock (Figure 108), and (5) dendritic (tributary) passages in the vicinity of the dolines and synclinal axis are parallel to the dip (Figures 108 and 109).

Detailed mapping and interpretation of these relationships have clarified the effects of lithology, structure, topography, and hydrogeology on the origin of the cave, as described below.

Lithologic Control

A comparison of the positions of passages in the cave with the stratigraphic section demonstrates the effects of lithology on speleogenesis in the Burnet area (Figure 88). Solutional enlargement has been greatest within the beds of limestone of the calcitic facies of the Gorman Formation. The master conduit extending from the Sam Bass Entrance to the Hall of Marble was largely excavated from a massive, 8.8 m-thick unit of limestone of the Gorman Formation, located 7.0 to 15.8 m below the Gorman-Honeycut contact (Figure 88). This unit is aphanitic, with massive beds (up to 0.9 or 1.2 m in thickness) alternating with strata composed of thin beds (2.5 to 5.0 cm in thickness). Passages attain their greatest cross-sectional dimensions in this unit, but in many places they extend upward into a fine-grained, thin-bedded unit of limestone (Figure 110). In the southern third of the cave most passages lie just above a 10.5 m-thick, microgranular, thin-bedded dolostone which is relatively resistant to dissolution (Figure 88 and 109). However, in places, pits or inclined passages have penetrated this unit and now communicate with lower levels developed within other beds of limestone (e.g. the pits east of the Cathedral Room which drop into the Lake Room Tunnel, shown in Figure 110, or the descending slope north of the Hall of Marble leading to the lower-level conduit at Catfish Lake). The Lake Room Tunnel and northern two-thirds of the cave have developed within aphanitic, medium-bedded beds of limestone (Figures 88 and 109). These passages, like the master conduit, have apparently been confined by overlying and underlying beds of dolostone of markedly lower solubility than that of the intervening beds of limestone.

Vertical confinement and perching of passages in Longhorn Cavern occur for the same reasons as described for the Bend area (Chapter 8, particularly Figure 92). Limestone, despite having coarser grain-size and lower primary porosity than dolostone (Goldich and Parmelee, 1947), is more conducive to forming caves because it is uniformly high in calcitic content (93.9 to 99.6 percent compared to 2.7 to 20.7 percent for dolostone). As in the Bend area, dissolution of limestone of low primary porosity has been promoted along avenues of secondary porosity (fractures or zones of collapse). Other lithic attributes, such as the presence of bedding-plane partings and the thickness of beds, are generally subordinate to the amount of calcite and degree of vertical fracturing in promoting dissolution during speleogenesis.

The lower dolomitic beds of the Honeycut Formation form a caprock over the upper calcitic beds of the Gorman Formation. The central third of the cave lies beneath the Honeycut; however, the northern and southern thirds lie beneath the uncapped outcrop of the Gorman Formation, just east of the exposed Gorman-Honeycut contact (Figures 105, 108, and 109). Extensive, solutional dolines occur within the outcrop of the upper Gorman but are markedly absent in the outcrop of the Honeycut to the west (Figures 105 and 108). The highly calcitic beds of the upper Gorman have promoted rapid infiltration and karstification, whereas the dolomitic beds of Honeycut and lower Gorman have not. The topographic surface to the northwest of the cave, along the synclinal axis, is nearly parallel to the dip (Figure 109), suggesting that a dip-slope has formed on the resistant Honeycut Formation.

Structural Control

Passages in Longhorn Cavern have been guided by both bedding attitude and fractures during speleogenesis. These structures are products of block-faulting which produced the Backbone Ridge Graben and the northwest-plunging syncline.

Relationship to Attitude of Bed. Longhorn Cavern trends generally parallel to the strike of the bedrock from its entrance near the synclinal axis to the siphon at the northern end of the cave along the synclinal flank (Figures 108 and 111). This implies that excavation of the cave occurred along horizontal levels within favorable (i.e. relatively soluble), stratigraphic intervals. This happens as conduits develop along flowpaths in the shallow-phreatic zone where the piezometric surface is either horizontal or of very gentle gradient.

Side-passages comprising the east side of the southern third of the cave are oriented down the dip and occur in close proximity to the synclinal axis. They appear to be aligned with the cluster of large dolines just to the east along the synclinal trough (Figures 108 and 111); and they apparently functioned as dip-oriented tributaries transmitting infiltrated water from the dolines to the strike-oriented, master conduit (Figure 109). The syncline may have served as a trough, guiding surficial runoff from a gently sloping, upland surface into the subsurface through the outcrop of the calcitic upper Gorman Formation. Vertical infiltration was arrested at the contact with the underlying dolomitic units, and this perched flow became further concentrated in the subsurface as it migrated down the plunging syncline (Figure 111).

Flow along the master conduit is assumed to have continued northwestwardly along the strike and within the calcitic beds of the Gorman, following the northeastern flank of the syncline. Discharge issued from one or more springs within the Roaring Spring Fault Zone (Figure 111). Which, if any, of the present springs along that trend may have served in this capacity has not been determined.

Relationship to Fractures. Orientations of linear segments of passages in Longhorn Cavern may be compared with orientations of major faults, mapped joints, and mapped photolineaments (Figure 112). Presumably, most fractures on Backbone Ridge were produced during the Pennsylvanian Period by regional tension, block-faulting, and concomitant synclinal folding adjacent to the Roaring Spring Fault Zone. Many fractures may have been subsequently widened or extended during release of confining pressure (unloading) as strata overlying the Ellenburger beds were removed by Cenozoic erosion.

Trends of fractures (Figure 112) and tectonic stresses may be summarized as follows (Figure 113):

1. A primary set of fractures (Set I) trends N 30°-60° E and includes fractures parallel to the Roaring Spring and Bald Knob Fault Zones. Fractures along this trend are more numerous and generally longer than those of other trends (Boyer and McQueen, 1964, figures 2, 3, and 4, p. 632-634).

Figure 110: Cathedral Room in Longhorn Cavern. Note elliptical cross-section of passage developed along a contact in upper beds of the Gorman Formation. Massive unit below contact is the major cave-forming unit. Beds dip 3 degrees to left (west). Pits at right drop into Lake Room Tunnel The passage floor has been entrenched by subsequent, vadose streams. The Frozen Waterfall is the massive deposit of flowstone along the right wall in background. View is to the north.



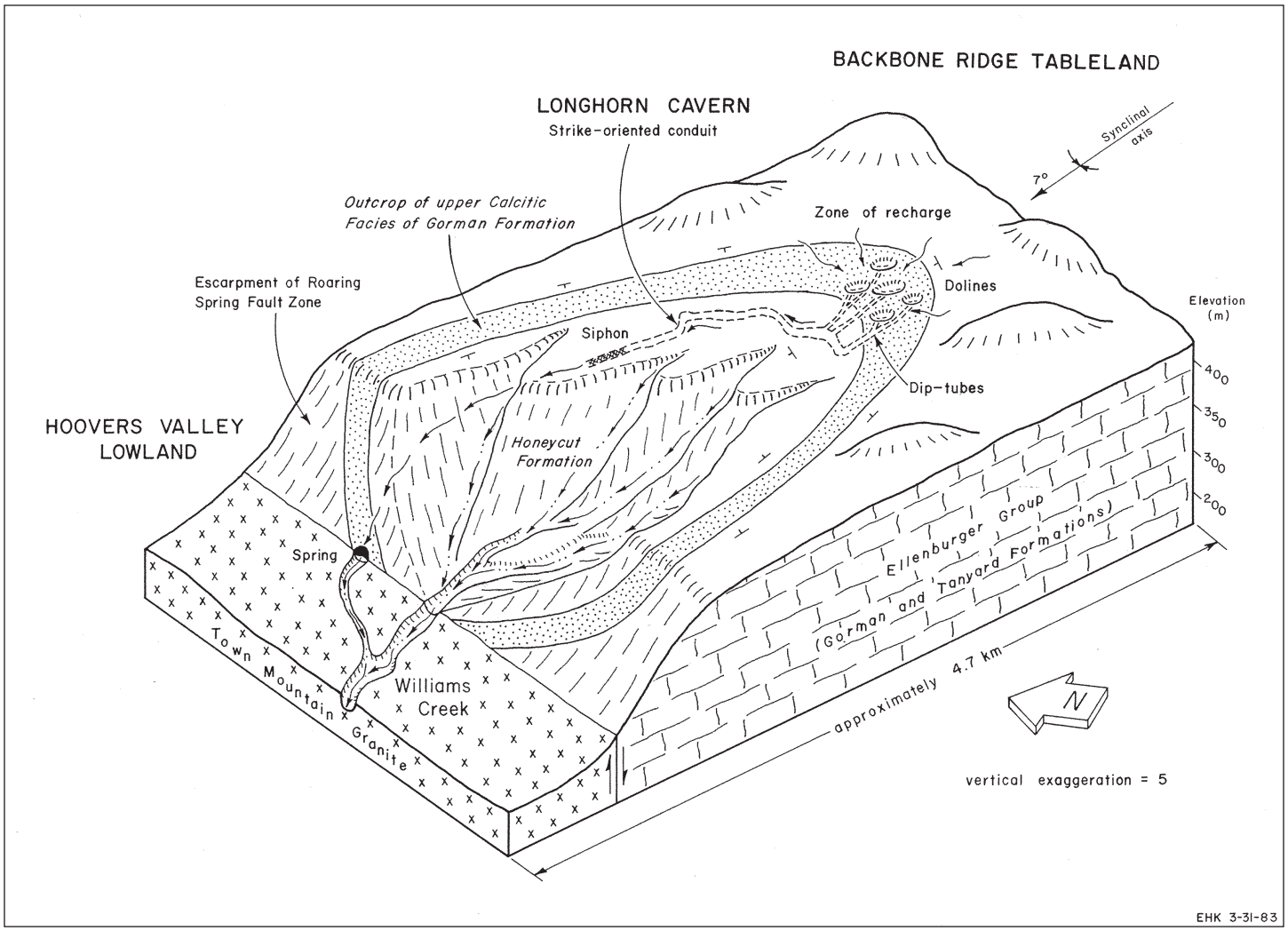


Figure 111: Generalized block-diagram of the vicinity of Longhorn Cavern showing structure and topography. Note Longhorn Cavern and its drainage lie within the upper, calcitic facies of the Gorman Formation (see also Figure 114)

2. A secondary set of fractures (Set II) trends N 30°-60° W. It includes many fractures that are parallel to the synclinal axis (N 60° W) which is perpendicular to the strike of the Roaring Spring Fault Zone. Fractures of Set II are well developed, but are somewhat shorter and less numerous than those of Set I (Boyer and McQueen, 1964, figures 2 and 3, p. 632-633).

3. Two conjugate sets of fractures (Sets IIIA and IIIB) trend N 0°-10° E and N 70°-80° E respectively. These form dihedral angles with fractures of Sets I and II and are significantly shorter than fractures of either of those sets (Boyer and McQueen, 1964, figures 2 and 3, p. 632-633).

4. Fractures of Sets I and II lie parallel to boundaries of fault-blocks and are perpendicular to the regional, northwestern-southeastern, tensional stress incurred during Pennsylvanian epeirogeny in the Llano region. Contemporaneous, northeastern-southwestern, compressional stresses may have occurred then as well (Figure 113), although these may have been slight and caused little deformation.

5. Conjugate Sets IIIA and IIIB are apparently products of simultaneous tension and compression (Figure 113). Therefore, fractures of all sets may be explained by a single tectonic episode.

6. It is interesting to compare the tectonic setting of the Burnet area (Figure 113) with that of the Langtry area (Chapter 5, Figure 28). Each region underwent northwestern-southeastern tension and northeastern-southwestern compression. Fractures group into sets that are both parallel to and perpendicular to these directions of stress and into conjugate sets at dihedral angles to these directions. These configurations are consistent with structural studies by Parker (1942), Muehlberger (1961), and Leonard (1977a).

Passages in Longhorn Cavern have developed along fractures of all four sets (compare Figures 112A and 113), but most segments of passages are aligned along the principal fractures of Set I. Presumably, these fractures are relatively open due to their tensional origin. Some other segments of passages follow open fractures oriented parallel to the synclinal axis.

A disproportionate number of segments of passages (in comparison to numbers of mapped joints and photolineaments) follow fractures of Sets IIIA and IIIB (compare Figures 112A,B,C and 113). This is easily explained by the arrangement of the fractures with respect to the attitude of the beds. Passages oriented along the north-trending fractures of Set IIIA include those of the main conduit that follow the strike of the bedrock in the vicinity of the nose of the syncline (Figure 108). These conduits were initiated in the shallow-phreatic zone near the intersection of the piezometric surface and the most soluble beds of the upper Gorman Formation (Figures 111 and 114). Passages along east-northeast-trending fractures of Set IIIB include those of the dip-oriented, eastern, tributary-passages in the southern third of the cave. These conduits conducted groundwater down the dip from the zone of recharge to the master conduit (Figures 111 and 114).

In summary, enlargement of passages has predominantly occurred along fractures of Sets I, II, and III. However, fractures fortuitously aligned along prevailing strike- and dip-oriented flowpaths have been selectively enlarged in greater proportion to those aligned obliquely to these flowpaths.

Modification by Collapse. Areas of collapse in Longhorn Cavern are few and are associated with development of dolines in the vicinity of the southern third of the cave. The Sam Bass Entrance is in a collapsed doline where the uppermost beds of the Gorman Formation have fallen into the cave. Collapse occurs where surficial denudation and dissolutional enlargement of dolines progress over pre-existing passages and chambers of caves. The thickness of the interval of limestone forming the roof of a cave is reduced to a critical dimension which is no longer self-supporting. The uppermost beds of the Gorman are thin-bedded and collapse readily into underlying voids. Ceilings of passages are weakest where conduits have attained large, cross-sectional dimensions or where high domes have developed. In places progressive stoping ultimately resulted in collapse of floors of dolines into the cave.

Presently a narrow and thin remnant of the uppermost beds forms a natural span over the Sam Bass Entrance (Matthews, 1963, figure 13, p. 18). A similar zone of collapse characterizes the former Main Entrance at the Dining Room.

Where strata overlying passages are thicker and stronger, collapse has not occurred. This is particularly true over those parts of the cave capped by the Honeycut Formation.

The mechanisms of stoping in passages and domes of Longhorn Cavern and concomitant collapse are very similar to those in Natural Bridge Caverns in the New Braunfels area. These phenomena are described in further detail in Chapter 10.

Hydrogeologic and Topographic Controls

The impetus for speleogenesis on Backbone Ridge was provided by block-faulting and subsequent inversion of the topography through erosion (Figure 106). Grabens of Paleozoic strata, including the Ellenburger Group, were faulted down and against surrounding Precambrian rocks of the basement during the Pennsylvanian (Figure 106B). The faulted topography was later planed by subaerial erosion in the Permian, Triassic, and Jurassic, and covered by shallow-marine sediments during the Cretaceous.

During regional denudation in the late Paleozoic and early Mesozoic the fault-blocks of the Llano Basin were eroded nonuniformly (Figure 106C). In some places exposed Precambrian granite and schist were less resistant to physical weathering and erosion than were the dense carbonate units of the Ellenburger Group. Although grabens, such as that of Backbone Ridge, were structurally lower than the surrounding basement-rocks, they remained topographically higher following differential erosion (Paige, 1912a; Stenzel, 1937). This topography was later buried under marine sediments during the Cretaceous (Figure 106D).

Renewed denudation and dissection of the Edwards Plateau and the Llano region began in the late Tertiary. This was in response to an increase in relief between the plateau surface and lowlands to the southeast created by Balcones faulting in the early Miocene. At first an erosional surface developed over central Texas by the mid-Pleistocene (Figure 106E). This was followed by rejuvenation, incision of the Colorado River and its tributaries, and re-exposure of the pre-Cretaceous topography which had previously developed on the Paleozoic and Precambrian rocks (Figure 106F). Backbone Ridge and many other nearby erosional outliers are relict and exhumed landforms. Their topographically flat uplands may represent former surfaces of planation and unconformable contacts between the Paleozoic and the Cretaceous (Stenzel, 1937).

Flow of groundwater in the Backbone Ridge Graben accelerated during the late Pleistocene as the surrounding topography was lowered with respect to the top of the ridge. Increased circulation along open fractures and within the soluble beds of the upper Gorman Formation promoted the development of caves (Figure 106G).

Phreatic Phase in Development of the Cave. The master conduit extending the full length of Longhorn Cavern originated within the uppermost few meters of the phreatic zone (Figure 114). Evidence for this includes (1) elliptical, cross-sectional shape of the passage (Figure 110), (2) strong concordance of the passage with bedding throughout most of its length, (3) orientation of the passage along the strike, (4) slightly undular profiles of the ceilings, and (5) low gradient of the passage.

Initial development of the master conduit began along prominent bedding-plane partings and along strike-oriented fractures. The peizometric surface remained relatively fixed for a considerable time during excavation of the master conduit. This may represent a hiatus in erosion of the Precambrian rocks adjacent to the graben. Lower-level, phreatic passages, such as the Lake Room Tunnel and the Wiggleys, may have evolved as the piezometric surface dropped in response to lowering of baselevel northwest of Backbone Ridge. Vadose flow through the master conduit had apparently been pirated to the Lake Room Tunnel through a series of pits in the vicinity of the Cathedral Room (Figure 110). The Wiggleys, which lies at a lower level and west of the master conduit, may have formed as a route of piracy and cutoff when the piezometric surface dropped. It is positioned slightly down the dip in the synclinal trough but remains within the calcitic facies of the Gorman Formation.

Vadose Phase in Development of the Cave. The tributaries leading from the dolines and joining the master conduit from the east, developed largely within the vadose zone (Figure 114). Evidence for this includes (1) canyon-like passages (narrow and high), (2) entrenchment of floors of passages, (3) orientation of passages down the dip, and (4) gradients of passages which closely match local angles of dip.

Tributary-passages formed where recharge entering dolines within the outcrop of the calcitic facies of the Gormam Formation was conveyed to the master conduit just below the "water table" (Figures 109 and 114). Flowpaths followed dip-oriented fractures that were generally aligned along the steepest gradients.

Flow of water in the vadose zone is purely gravitational, moving down the

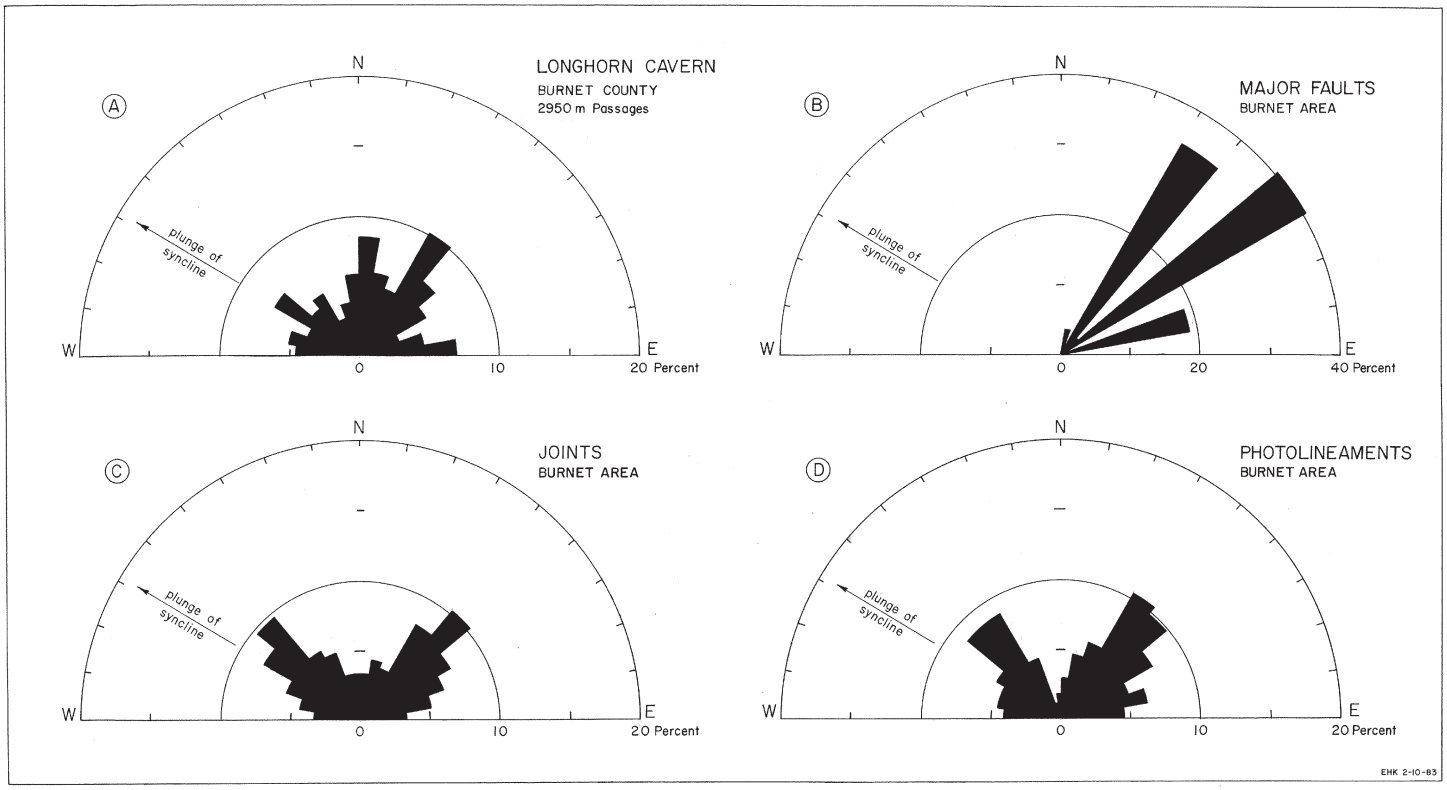


Figure 112: Rose-diagrams for the Burnet area showing: (A) Linear segments of passages in Longhorn Cavern, (B) Major faults of the Burnet area, (C) Joints in Ellenburger rocks mapped by Boyer and McQueen (1964, figure 6A, p. 634), (D) Photolineaments in Ellenburger rocks mapped by Boyer and McQueen (1964, figure 6C, p. 634). Compare with Figure 113.

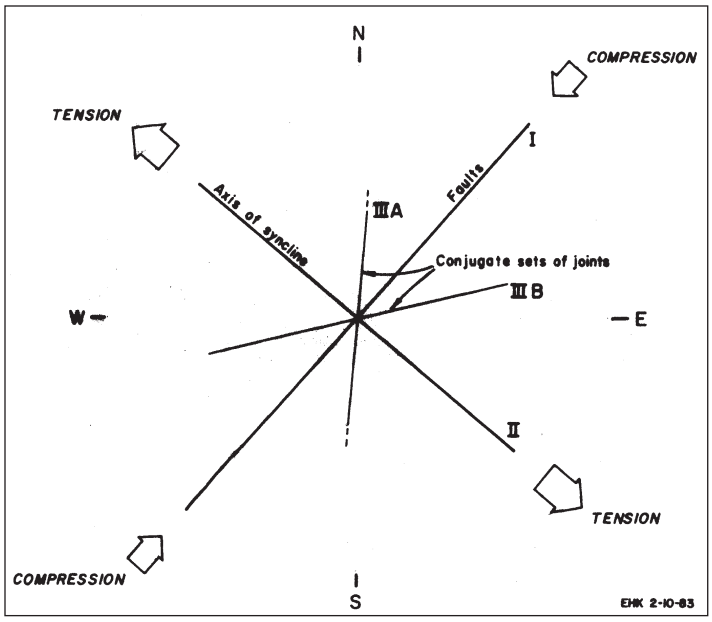
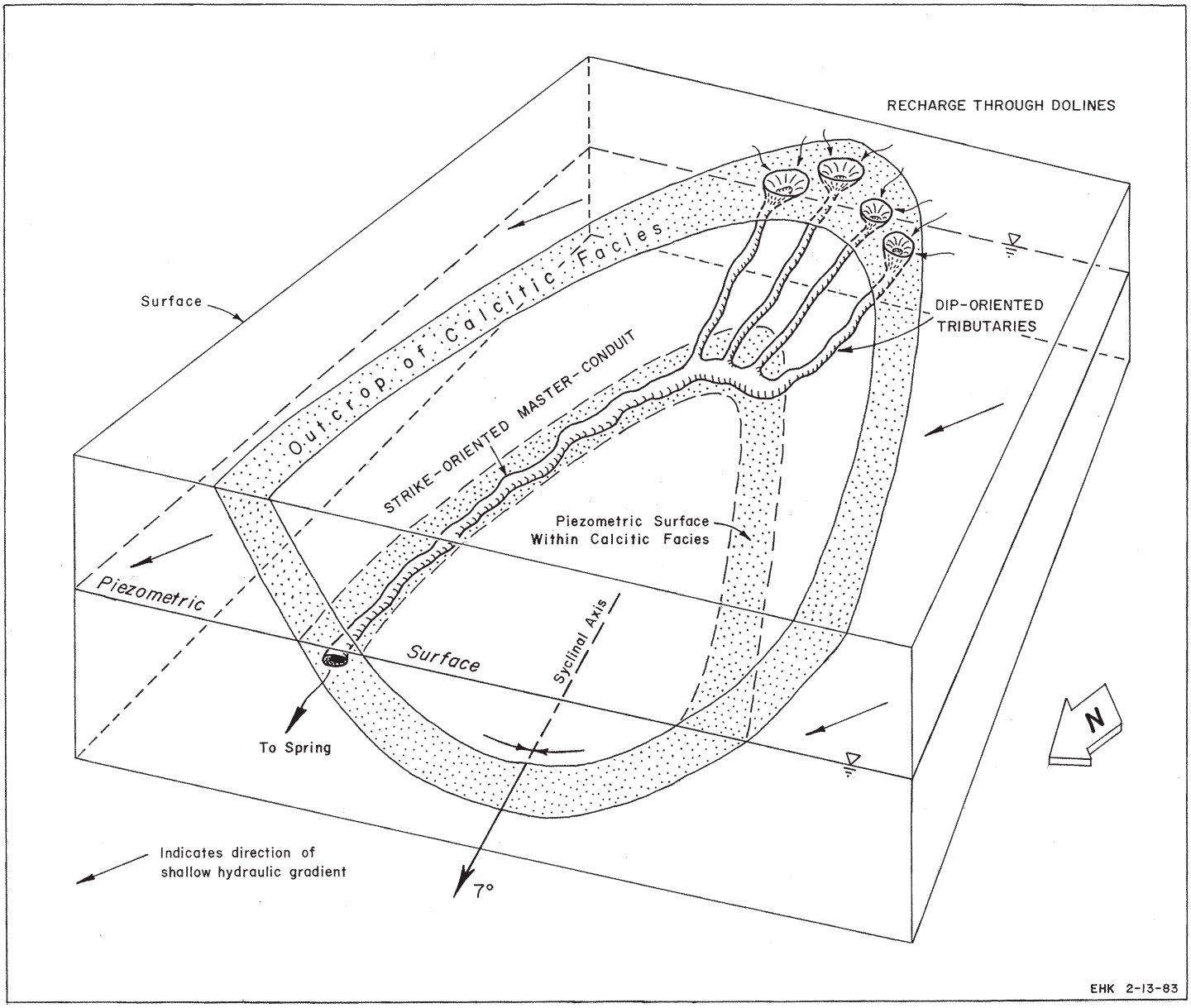


Figure 113: Schematic summary of major trends of fractures, regional stresses, and structural elements in the Burnet area. Set I fractures are of the primary set, parallel to faults. Set II fractures are of the secondary set, perpendicular to faults and parallel to synclinal axis. Set III fractures are conjugate sets at dihedral angles to Sets I and II. Compare with Figure 112.

Figure 114: Generalized schematic block-diagram showing origin of Longhorn Cavern. Strike-oriented, master conduit has developed along the intersection of shallow-phreatic zone and calcitic facies of the Gorman Formation. Dip-oriented, vadose tributaries have developed along steepest gradients from zone of recharge at dolines to master-conduit. Drawn with vertical exaggeration and not to scale; compare with Figures 109 and 111).



steepest, available flowpaths. This is in contrast to flow in the phreatic zone which is forced to move along the steepest hydraulic gradients. Palmer (1977) has noted distinct differences in morphology between tributary passages formed in the vadose zone and master conduits formed in the phreatic zone just below the "water table" (piezometric surface). The change in morphology is easily detected at junctures of the two types of passages, in Longhorn Cavern as well as other caves in general. There is a transition from the canyon-like cross-section of the "vadose passage" to a tubular or elliptical cross-section of the "phreatic conduit." Usually, vadose passages oriented down the dip become phreatic conduits oriented along the strike, and there is an accompanying reduction in gradient of the passage. The relationship of speleogenesis to structure in Longhorn Cavern is similar to that along Cleaveland Avenue in Mammoth Cave, Kentucky (Palmer, 1977, 1981a, p. 99-103). In both caves, phreatic, master conduits follow the strike around noses of gentle synclines, and vadose tributaries trend down the dip along the synclinal troughs.

As the piezometric surface dropped beneath the master conduit in Longhorn Caverns, flow contributed by the tributary-passages and originating as continual recharge at the dolines, was transmitted northward along the master conduit as free-surface flow. During this stage the entire cave functioned as a system of vadose flow, except during intervals of flooding (see below). Channels cut into the floor of bedrock of the master conduit (Figure 110; see also Matthews, 1963, figures 36 and 37, p. 42-43) and thick sequences of fill (clay and silt) throughout the cave are evidence of long periods of vadose discharge. Semken (1960b, 1961) recognized four stratigraphic units of fill in the cave. A red fill, the most extensive deposit in the cave, is at least Late Pleistocene (Wisconsinan), based on the remains of vertebrate fauna it contains, but it may represent Kansan terra rosa washed into the cave during the Late Pleistocene. A less-extensive, overlying, black fill is considered to be Holocene (Frank, 1964, p. 4-6).

Dolines on the surface presently contribute great amounts of recharge during storms. Bottoms of dolines are highly karsted, and many deep (2 to 5 m) fissures, similar to those of the Bend area (Chapter 8), are open to the surface. The dolines drain approximately 1.0 km^2 of the tableland. The largest depression is that containing the Sam Bass Entrance, and its large volume of recharge relative to other dolines may account for the relatively large dimensions of the dip-oriented tributary leading from it.

Modification by Floodwater. Longhorn Cavern has been modified by periodic flooding. The main conduit and several tributaries exhibit properties produced by incursion of floodwater, including (1) passage-spurs that abruptly end in bedrock (Figure 107), (2) abrupt changes in cross-section of passages, (3) naturally excavated banks of sediment, (4) maze-like configurations of passages, and (5) pockets and bell holes in ceilings (see photographs in Matthews, 1963, figures 14, 22, 30, 32, 38).

Apparently vadose recharge periodically ponded in the upstream parts of the cave, behind constrictions along the master conduit. The primary constriction was at Catfish Lake, where dimensions of passages were markedly reduced. Periodic flooding was probably most intense during pluvial periods when local climate was cooler and wetter. The chemical aggressiveness of the floodwaters is indicated by redissolved speleothems throughout most of the cave. The best example, the Rock of a Million Layers (Matthews, 1963, figure 19, p. 23), occurs along an inclined segment of the master conduit, just upstream from the constriction at Catfish Lake. Redissolution had been intensified at this location by a relatively steep, hydraulic gradient. Catfish Lake and other parts of the cave still undergo occasional flooding during large storms.

Chronology of Speleogenesis

The following chronological sequence is proposed for the origin of Longhorn Cavern:

1. Paleozoic strata, including those of the Ellenburger Group, were deposited unconformably on the Precambrian basement (Figure 106A). Tensional stress and block-faulting accompanying uplift of the Llano region during the Pennsylvanian produced many down-faulted blocks (grabens) in the Llano region. In places Paleozoic strata were juxtaposed against crystalline, Precambrian rocks (Figure 106B).

2. During faulting the Backbone Ridge Graben was folded into a gentle syncline with an axis trending perpendicular to its border-faults (Figures 108, 109, 111, and 114). This activity produced systematic fractures consisting of two principal sets trending parallel and perpendicular to the faults and two conjugate sets forming dihedral angles with the other sets (Figure 113).

3. The region was eroded during the late Paleozoic and early Mesozoic, resulting in an inverted topography, owing to differential erosion of the crystalline rocks of the basement with respect to the Paleozoic carbonate rocks (Figure 106C). Shallow-marine, carbonate sediments were deposited over the area during the Cretaceous (Figure 106D).

4. Rapid erosion commenced in the late Tertiary, following uplift of the Edwards Plateau and Llano region during movement along the Balcones Fault Zone. Central Texas was planed to a subdued topography by the mid-Pleistocene (Figure 106E). Ensuing entrenchment of the Colorado River and an increase in local, topographic gradients promoted rapid removal of the Cretaceous cover from much of the Llano region.

5. Continued erosion of the Llano Basin during the Pleistocene re-exposed the inverted topography of the pre-Cretaceous (Figure 106F). This resulted in the present setting, where structurally lower Paleozoic rocks of the graben form topographical plateaus or outliers within the evolving Llano Basin.

6. Increase in local relief in the vicinity of Backbone Ridge promoted the circulation of groundwater and the onset of speleogenesis (Figure 106G).

7. The piezometric surface dropped in response to lowering of baselevel. The master conduit of Longhorn Cavern developed in the shallow-phreatic zone and along the strike in the relatively soluble, calcitic facies of the Gorman Formation. The development of passages was guided by strike-oriented fractures and along prominent bedding-plane partings. Flow was to the north and northwest, following the northeastern flank of the syncline and a shallow, northwesterly hydraulic gradient (Figure 114). Concurrently, recharge became concentrated within the outcrop of the calcitic facies along the axis of the syncline (Figure 111). Solutional dolines developed within this area of the tableland, and infiltration through these was transmitted through an evolving series of tributaries to the master conduit that was oriented down the dip within the vadose zone.

8. A later lowering of baselevel and of the piezometric surface resulted in vadose conditions throughout the cave. Most water was conducted to lower-level conduits, which were evolving in the new, shallow-phreatic zone.

9. As the piezometric surface continued to drop during the Pleistocene, the cave was periodically subjected to flooding, especially during colder and wetter, pluvial climates. This resulted in modification of the dimensions of passages, in domed ceilings, in numerous joint spurs, and in redissolution of speleothems.

10. Great amounts of detrital sediments were deposited in the cave by vadose waters during the Wisconsinan glacial stage. Some of these were later re-excavated by floodwater; but, in some places, sediments have armored floors of bedrock and have prevented subsequent entrenchment of vadose streams.

11. Continued denudation of the surface caused ceilings of the cave to collapse where cross-sections of the passages were large and nearer to the surface. This created entrances at the bottoms of several dolines.

SUMMARY

Speleogenesis in the Burnet Area has been confined to a topographically high outlier of carbonate rock. Passages in Longhorn Cavern consist of two general types: (1) shallow-phreatic conduits developed within soluble beds and along the strike of the strata, following strike-oriented fractures and bedding-plane partings, and (2) vadose tributaries at the headwaters of the cave that drain a karsted tableland and are oriented down the dip, following dip-oriented fractures. In many respects Longhorn Cavern is the best example of a fully integrated cave-system on the Edwards Plateau. Water from discrete points of recharge has been concentrated by a dendritic system of vadose feeders and then transmitted along a master, shallow-phreatic conduit to springs at baselevel. Concordance of conduits with the strata and structural control of passages by attitude and fractures have been very well maintained during speleogenesis.

10
NEW BRAUNFELS AREA

DESCRIPTION OF AREA

The New Braunfels area lies within Comal, Bexar, and Guadalupe Counties (Plate 1). It is bounded on the south and north by lines of 29°35'00" and 29°50'00" north latitude respectively and on the east and west by lines of 98°02'30" and 98°22'30" west longitude respectively. The northeastern part of the area is drained by the Guadalupe River and the southeastern part by Cibolo Creek, a tributary of the San Antonio River (Plate 2).

The geology of the area has been largely mapped at scales ranging from 1:24,000 to 1:250,000 (George and others, 1947, 1952; Newcomb, 1971; Barnes, 1974b). Newcomb (1971) described the geology of the Bat Cave 7.5-minute U.S. Geological Survey quadrangle which encompasses the major caves described in this chapter.

The New Braunfels area is generally underlain by Lower and Upper Cretaceous rocks (Figure 115). The lowest cavernous unit is the upper member of the Glen Rose Formation, a generally thin-bedded sequence of fine-grained dolomite and marl of variable hardness (Stricklin and others, 1971). Beds are alternatingly resistant and recessive and weather to form stairstep-topography. The uppermost beds are massive and dolomitic. Up to 17 m of the upper member of the Glen Rose are exposed along valleys of incised streams.

The Glen Rose is overlain by the Bull Creek Limestone and Bee Cave Marl Members of the Walnut Formation. The Bull Creek is a thin- to medium-bedded, hard, well-sorted biomicrite, intramicrite, and fossiliferous intrasparite, and the Bee Cave is a clayey, fossiliferous micrite and fossiliferous marl (Moore, 1959, 1961a,b, 1964). The Walnut is approximately 14 m thick in this area.

The Kainer Formation of the Edwards Group overlies the Walnut and consists of aphanitic to coarse-grained, massive- to thin-bedded, hard, brittle, resistant limestone. It is locally dolomitic and bored, contains abundant chert, and in its basal part is nodular, honeycombed, and contains zones of solutional and collapsed breccia.

Younger rocks of the Lower and Upper Cretaceous crop out in much of the southeastern part of the area, within and to the southeast of the Balcones Fault Zone. Deposits of Quaternary alluvium are abundant in valleys of larger streams, particularly along Cibolo Creek (Figure 115).

The New Braunfels area overlies the faulted outcrop of the Edwards Group (Plate 3), bounded on the northwest by the Tom Creek Fault and on the southeast by the Comal Springs Fault (George and others, 1947, 1952, plate 2; Kastning, 1978, plate 1). The outcrop of the Edwards Group within the Balcones Fault Zone is 20 km wide in this vicinity, and recharge enters the unconfined zone of the Edwards Aquifer by direct infiltration. The bed of Cibolo Creek between Boerne and the western boundary of the outcrop of the Edwards is part of the zone of recharge for the Edwards Aquifer. Water is lost to the Glen Rose Formation along the bed of Cibolo Creek, and is transmitted into the Edwards Limestone across the Hidden Valley and Bear Creek Faults, where the Glen Rose and Edwards are in juxtaposition (see profiles of George and others, 1947, 1952, plate 2, and Kastning, 1978, plate 1). Comal Springs (Figure 14), within the New Braunfels city limits, is the largest exsurgence of groundwater from the Edwards Aquifer (Guyton and Associates, 1979).

The New Braunfels area lies along the southwestern flank of the San Marcos Arch (Figure 10). The arch was raised along its northwest-trending axis during the close of the Early Cretaceous (Rose, 1968, 1972; Woodruff and Abbott, 1979a,b; see also Chapter 2). The deformation that produced the arch created several sets of fractures. Those of a northwestern orientation lie parallel to the axis of the arch and are accompanied by two sets of conjugate fractures. These sets resulted from extensional stresses during uplift. Later deformation during Balcones faulting produced another set of extensional fractures trending to the northeast and parallel to the individual faults (Kastning, 1981b, 1984).

Subaerial exposure of Lower Cretaceous rocks brought about by uplift of the San Marcos Arch and incision of streams draining the Edwards Plateau in response to Balcones faulting during the Miocene were crucial in the evolution of circulation of karstic groundwater and development of caves. The chronology of these events and the influence of geologic structure on speleogenesis are described in detail later in this chapter.

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

There are several large caves in this area, including the Natural Bridge Caverns System, Bracken Bat, Double Decker, Ebert, Dinosaur, Bear Creek, Beal Ranch, Rompel, Zuercher No. 1, and Brehmer Caves (Plate 1 and Appendix C). White (1948a) and Reddell (1964a,b) have described these caves. Beck (1968, 1973) mapped several caves in Comal County and classified them into three categories based on their relationship to the flow of groundwater: (1) "influent caves", which are generally vertically oriented and formed by water flowing from the surface, through dolines, to the water table, (2) "effluent caves", which are generally horizontally oriented and formed by water flowing from the aquifer to the surface, emerging as springflow, and (3) "conduit caves", which are formed by water flowing just beneath the water-table, without connection to the surface except by secondary modification. Beck's primary purpose was to relate speleogenesis in Comal County to the "classical" theories of the development of caves.

Knox (1975) inventoried points of recharge, including caves and enlarged fractures, in the upper basin of Cibolo Creek in Kendall, Comal and Bexar Counties. She correlated these with available hydrologic data for the basin.

Abbott (1973, 1975, 1977a,b) investigated the hydrogeology of the Edwards Aquifer in the vicinity of New Braunfels and attempted to relate development of caves to the diagenetic evolution of the Edwards Limestone and to the hydrochemistry of the aquifer. The results from this analysis were later incorporated with an interpretation of the evolution of the drainage-basin into a plausible hypothesis for the development of cavernous porosity in the area of the San Marcos Platform (Woodruff and Abbott, 1979a,b).

Geologic studies undertaken in recent years in the Natural Bridge Caverns System have focused on the influence of lithology and structure on the development of the cave (Kastning, 1978, 1980e, 1981b, 1984; Knox, 1979, 1981). Renewed surveying of the cave by Orion Knox and others has produced one of the most detailed maps of a major Texas cave (Kastning, 1978, plate 2 and Fieseler and others, 1978, p. 91). Beck (1978) investigated the coloration of flat-topped calcitic stalagmites in Natural Bridge Caverns known as "fried-egg" formations. He determined that yellow and white zonation of color results from differences in the scatter of light by two zones of calcite having disparate crystalline structure. Chemical analyses of water from Natural Bridge Caverns are described by Mulet and Sanchez (1973)

DEVELOPMENT OF CAVES

In the New Braunfels area this study focused on the genetic interpretation of three large caves located immediately to the northwest of the Bat Cave Fault, one of the major components of the Balcones Fault Zone. The Natural Bridge Caverns System, Bracken Bat Cave, and Double Decker Cave have formed in an area between this fault and a large meander in Cibolo Creek to the northwest (Figure 115).

Descriptions of Caves

Natural Bridge Caverns System

The Natural Bridge Caverns System, in the west-central part of the Bat Cave Quadrangle, consists of three separate caves: the North Caverns, the South Caverns, and the Jaremy Room (Figure 116). These are remnants of a once larger cave now segmented by collapse and development of dolines. Most known passages are shown on the map, but several side-passages at the north end of the caverns remain to be surveyed. Total surveyed length of the system is presently 3354 m. The North Caverns, containing the commercially developed passages, is presently the ninth longest cave in Texas (Appendix C). The end-to-end extent of Natural Bridge Caverns System is 1160 m. Detailed descriptions of the system, history of its exploration, and bibliography are given by Knox (1962), Reddell (1964a, p. 39-48; 1964b, p. 36-40), J. Black (1973), and Black and Clark (1973).

Chambers of the Natural Bridge Caverns System are as large as any known from caves of the eastern Edwards Plateau. Speleothems within the cave are among the largest and most spectacular in Texas (Figure 117).

The only natural entrance to the system is in a large collapsed doline directly behind the present visitor-center. A natural span of limestone bridges the doline, providing the namesake for the cave (Figure 118). The doline, entrance, and side-passages to the southwest have been known for some time.

Pluto's Anteroom was discovered in 1960, and all passages to the north were explored during the period 1960 to 1963. The North Caverns, from the entrance to the Hall of the Mountain Kings, was developed for tourism and opened to the public in 1964 (Heidemann, 1979). An artificial tunnel was excavated between the surface and the Hall of the Mountain Kings in order to provide an exit for tours. It has intersected a small, unnamed cave midway along its length.

The management of Natural Bridge Caverns suspected that a continuation of the large, main passage of the North Caverns might exist to the south, beyond the collapsed doline. This was confirmed by exploratory drilling in the late 1960's when five boreholes were driven to allow entry into the new chambers. One hole intersected the Jaremy Room and another pierced the South Caverns and a short, overlying passage. Views of the cave in plan and profile (Figure 116) clearly suggest that all segments of the system were once interconnected.

The plan-view of Natural Bridge Caverns (Figure 116) shows that most passages consist of linear segments oriented along several preferred directions. An overall north-south trend persists, particularly for the passages of larger cross-section. Several east-trending side-passages join the main cave.

The profile (Figure 116) clearly indicates that levels of passages have developed along several, favorable horizons. Vertical position of levels in the cave is correlative with stratigraphic position (Figure 119). The dip is subhorizontal (less than one half of a degree to the southeast) in this vicinity. Large segments of the master conduit on the lowest level are wholly within the Upper Glen Rose Formation, including the South Caverns, Purgatory Creek, Grendel's Canyon, and the Limbo and Lake Passages. Above this level is another significant horizon of passages which includes the South Fault Passage, Chapel Hall, Jan's Long Crawl, Emerald Lake Passage

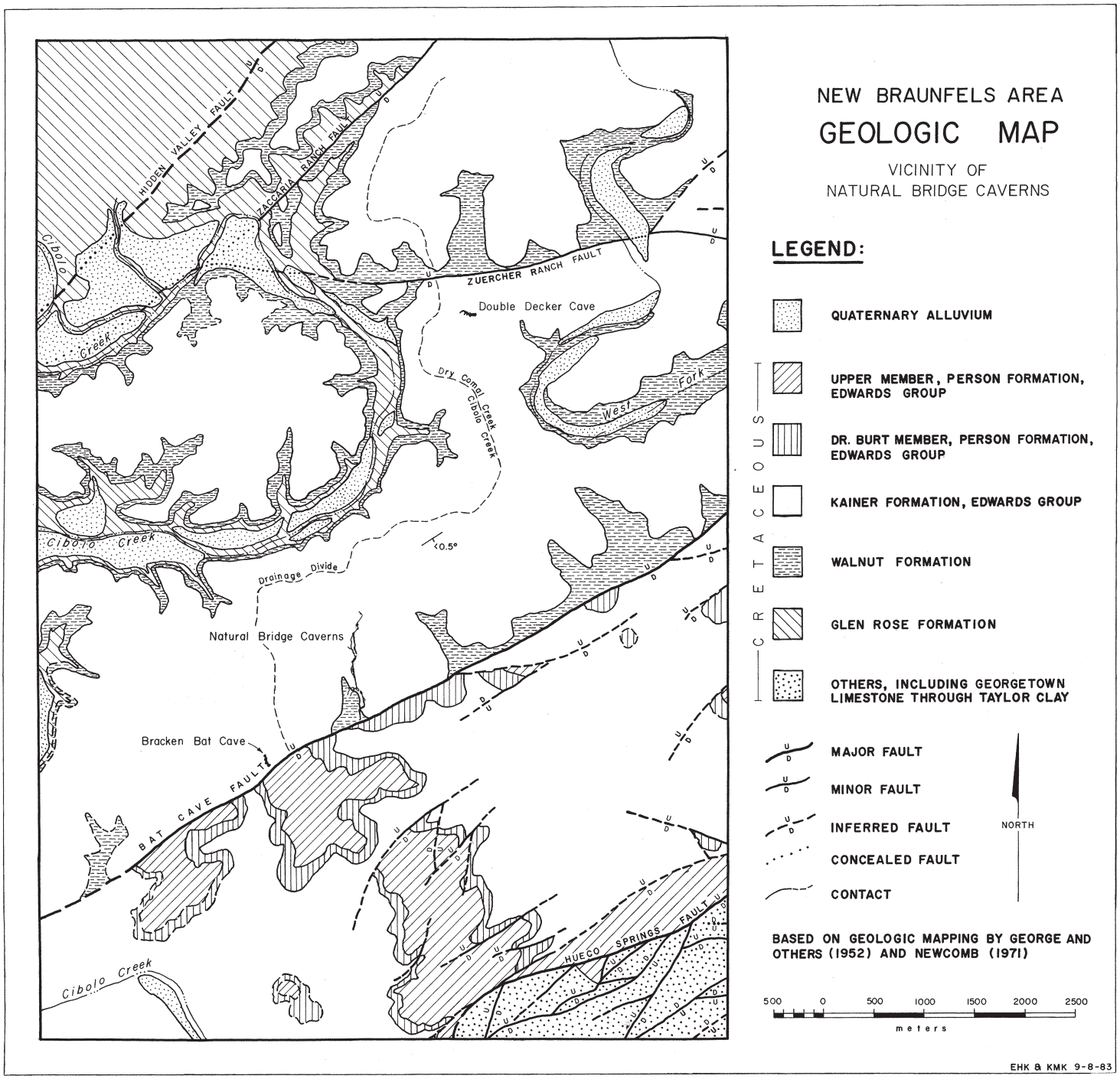


Figure 115: Geologic map of the New Braunfels area, Comal and Bexar Counties. Modified from George and others (1947, 1952) and Newcomb (1971, plate 1)



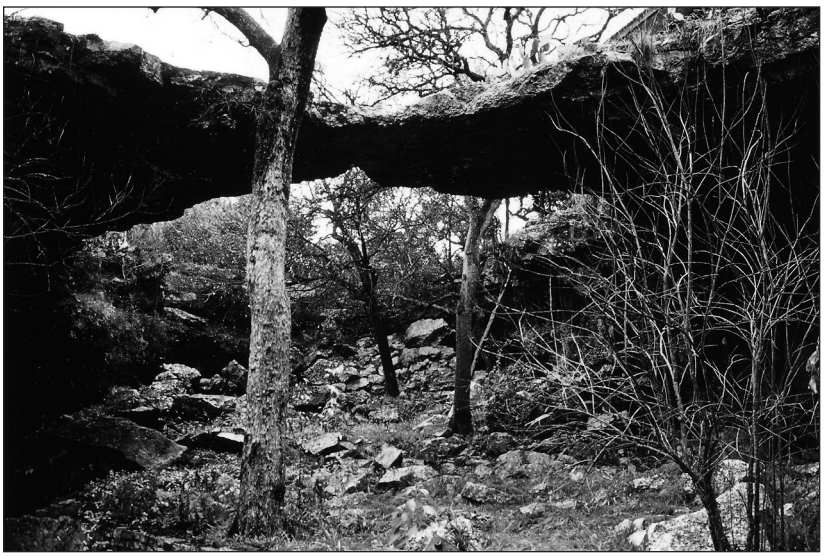
Figure 117: Stereoscopic photograph of the Mount of the Landlords in the Castle of the White Giants, Natural Bridge Caverns. Height is approximately 5 m. Niches were formed by periodic flooding. View is to the west.

and some shorter segments between the Inferno Room and Dome Pit. This level is also within the Upper Glen Rose, but nearly at its contact with the overlying Bull Creek Limestone Member of the Walnut Formation. Passages of shorter extent are found on two or more levels in the Walnut Formation. These are represented by (1) the isolated passage overlying the South Caverns, (2) St. Mary's Hall, (3) the Coon Rooms, and (4) the passage from the Moors to the Inferno Room and Dome Pit.

Natural Bridge Caverns presently contains two watercourses, both of very short extent within the cave. Purgatory Creek enters on the eastern wall of the main passage at one of the lowest points in the cave (64 m below the entrance). It curves once across the passage and leaves through the eastern wall only 18 m south of where it enters. These openings are alluviated and exploration is not possible. A second watercourse, River Styx, enters and leaves the northernmost section of the cave twice in its southward flow. River Styx may very well be an upstream segment of Purgatory Creek, but this remains to be confirmed by techniques of water-tracing. A well has been driven into the cave at River Styx from which water is pumped to the surface for use at the visitor-center.

Purgatory Creek is frequently dry, but during wet periods it increases in flow. Storms of high intensity cause sufficient flow in Purgatory Creek to inundate lower levels of the cave. A storm on May 11-12, 1972, caused devastating floods on the Guadalupe River and nearby Blieders and Dry Comal Creeks (Colwick and others, 1973; Baker, 1975, 1977). Three days after the storm water in the cave rose approximately 18 m above the bed of Purgatory Creek. After the crest of the flood the level of water began dropping at about 20 cm per day (Knox, 1981). The floor of the Castle of the White Giants had been under approximately 2 m of water at this time. This is the largest flood recorded in the cave since its discovery.

Figure 118: The Natural Bridge spanning the collapsed doline at the entrance to Natural Bridge Caverns. Excavated entrance is at lower left. View is to the north.



Water enters the cave by infiltration even during long, dry periods. Seepage is sufficient to promote continual deposition of speleothems in most passages. The Chandelier formation, a large drapery-stalactite in the Castle of the White Giants, normally receives enough flow to allow water to fall from it in a continuous trickle rather than in drops. Moisture and humidity are noticeably high throughout the cave. Glass doors have been installed in the entrance and exit tunnels to maintain moisture levels and sustain the growth of speleothems.

The Natural Bridge Caverns System is situated in the uplands above Cibolo Creek (Figure 115). The natural entrance is approximately 34 m above the bed of the creek and Purgatory Creek is approximately 26 m below it. The surface is drained to the southeast, away from Cibolo Creek and into Bear Creek, a tributary of Dry Comal Creek and the Guadalupe River.

Bracken Bat Cave

Bracken Bat Cave is 960 m to the southwest of Natural Bridge Caverns (Figure 115). The entrance is in a doline 30 m in diameter and 10 m deep. The cave extends N 18° W for 130 m as a single

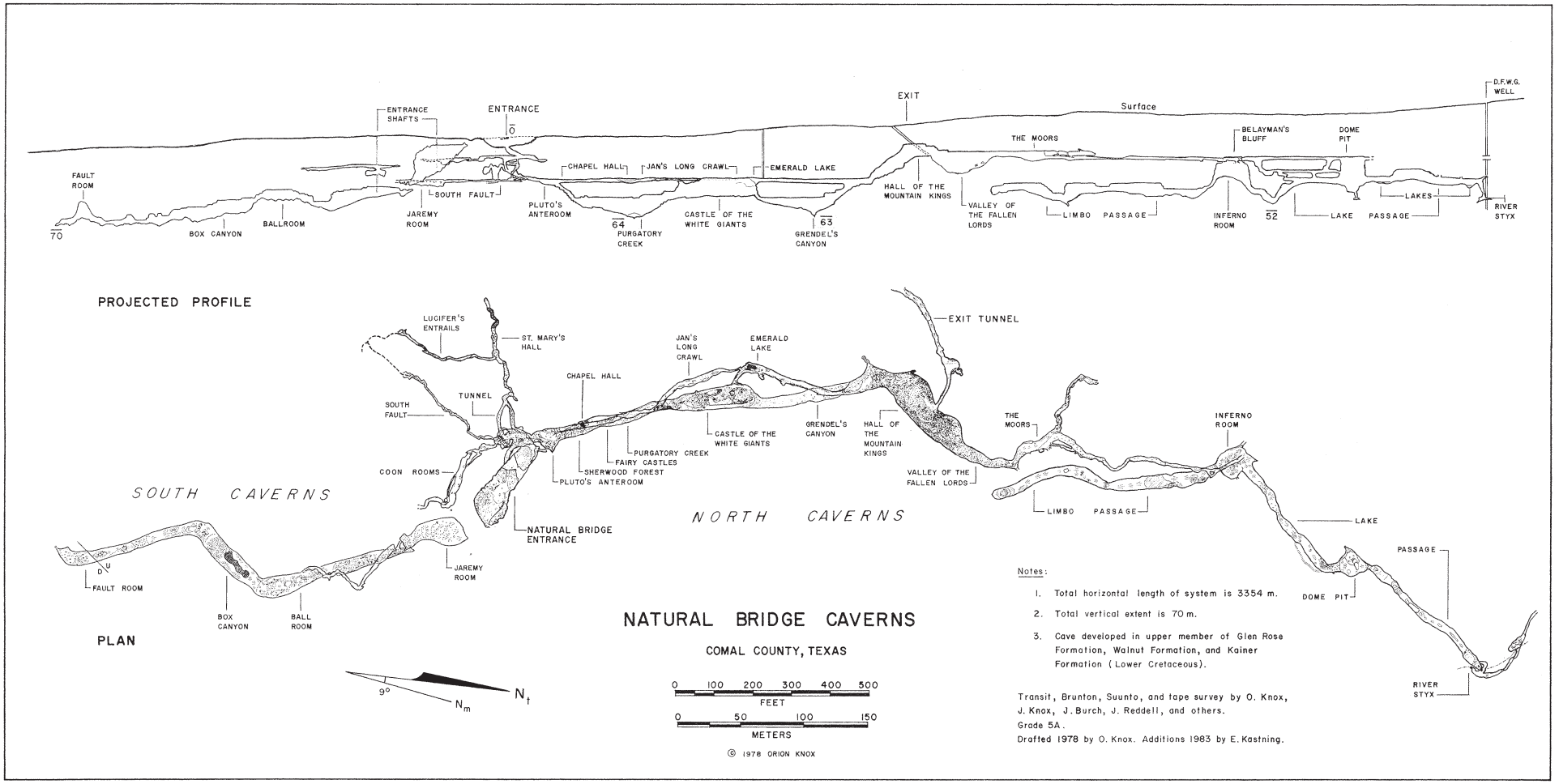


Figure 116: Map of the Natural Bridge Caverns System. Modified from Kastning (1978, plate 2) and Fieseler and others (1978, p. 91).

large passage, averaging 20 m in width and 10 m in height (Figure 120). From the entrance the passage descends at 35° as a debris-slope. About 30 m inside the cave and approximately 30 m below the surface the passage continues horizontally, and 105 m further it terminates in a large dome 25 m in height and 25 m in diameter. The floor of this chamber is 36 m beneath the surface, and there is a smaller dome, 21 m in height, on the northeastern wall. Severed by collapse of the doline at the entrance, the cave may extend to the northwest, but to date no such passage has been discovered.

Bracken Bat Cave has developed on one level. Stratigraphically, the main passage is in the Walnut Formation, just below its contact with the Kainer Formation of the Edwards Group (Figure 119). The slope of the entrance passes through the lower part of the Kainer Formation.

No active flow of water occurs in Bracken Bat Cave. Surficial runoff flows south directly to Cibolo Creek by way of a smaller, unnamed tributary. The drainage divide between Cibolo and Dry Comal Creeks passes between Bracken Bat Cave and the Natural Bridge Caverns System.

Bat Cave Fault, named after the cave, is approximately 200 m to the southeast. Terrain in the vicinity of the cave is moderately dissected.

Bracken Bat Cave contains one of the world's largest colonies of Mexican free-tailed bats, estimated at 20 million individuals. Guano has been sporadically mined from the cave since the Civil War. A 9 m-deep shaft was driven into the large dome at the rear of the cave to facilitate removal of guano. Further historical and biological information as well as bibliographic references may be found in Reddell (1964a,b) and Newman and Black (1973).

Double Decker Cave

Double Decker Cave is 3.93 km north-northeast of Natural Bridge Caverns in the northwest part of the Bat Cave Quadrangle (Figure 115). The cave has 208 m of surveyed passage and attains a maximum depth of 31.7 m below the entrance (Figure 121). A 14.3 m-deep pit at the base of a small doline, 7 m in diameter and 2 m deep, provides entrance to the cave. Although the cave has been known locally for several decades, it was first mapped and studied during field-work for this investigation.

Double Decker Cave has developed on two levels and along two major directions. The Upper Level extends N 30° W for 45 m from the entrance before turning to the southwest for 20 m. A 15 m-long passage, also trending southwest, extends from the Lower Level at the base of the slope between the two levels. Although Double Decker Cave is relatively short, its passages have large cross-sections. The Upper Level averages 10 m in width and 2 to 15 m. in height and the Lower Level averages 8 m. in width and 4 m in height. At the east end of the cave the Upper Level terminates in a large room, the Hall of Teetering Totems, which is 20 m in diameter and 12 m high. This room contains numerous large blocks of breakdown and tall (up to 3 m) but very narrow stalagmites known as "broomsticks" or "totems". One slender stalagmite, 2.4 m high, is only 4 cm in diameter. Many similar but smaller formations are found along the lower passages.

Areas of collapse occur in the vicinity of the entrance and at the beginning of the Upper Level. Moreover, some blocks of bedrock have slumped in the Hall of Teetering Totems. Much of the cave is floored by flowstone up to 10 cm thick. This is most prevalent around two large white columns in the Upper Level, in the Lilliputian Forest, on the steep slope between the two levels, and at the western end of the cave.

The pit at the entrance is in the Kainer Formation, the Upper Level has developed primarily in the Walnut Formation, and the Lower Level is wholly within the Glen Rose Formation (Figure 119). Floors of passages are generally covered by mud, breakdown, or flowstone.

Double Decker Cave does not contain an active stream. However, the western end of the cave receives a moderate amount of seepage through infiltration, accounting for the active deposition of flowstone, dripstone, and dams of travertine, and for the presence of small, standing pools of water.

Drainage in the vicinity of Double Decker Cave is to the southeast into the West Fork of Dry Comal Creek (basin of the Guadalupe River). The local terrain is moderately dissected.

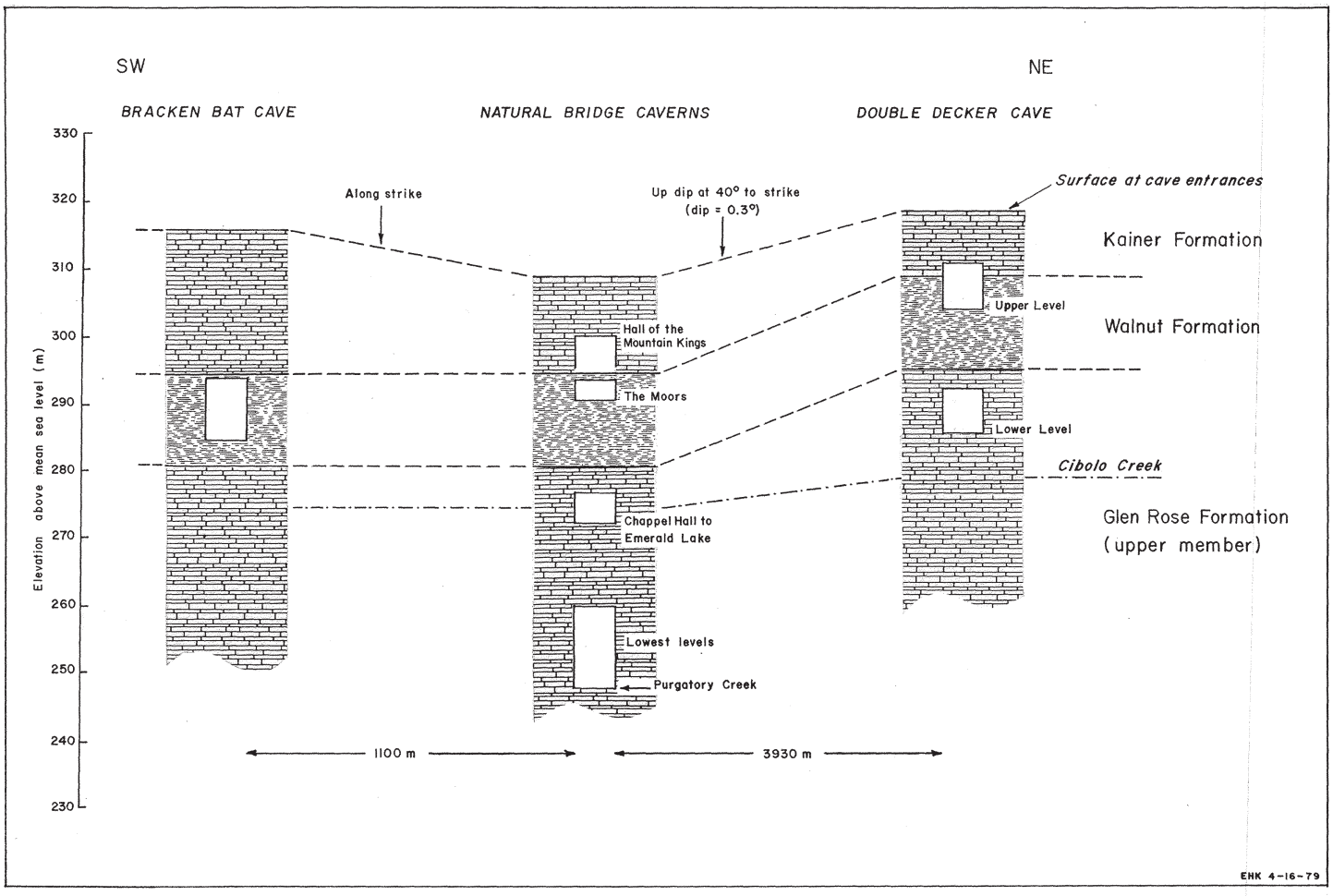


Figure 119: Correlation of horizons of passages within Natural Bridge Caverns System, Bracken Bat Cave, and Double Decker Cave. Topographic elevations at the entrances to the caves and at the nearest point on Cibolo Creek are indicated.

Speleogenesis

Many caves of the New Braunfels area, including some of the largest ones, have developed in the upper member of the Glen Rose Formation despite a much greater areal exposure of the Edwards Group than that of the Glen Rose. Dissolution and speleogenesis have occurred within the alternating sequence of calcareous, marly, and dolomitic beds of the Glen Rose, but relatively little development has taken place in the overlying, crystalline, Edwards Limestone. Reddell (1964a) and Beck (1968) have commented on this apparent anomaly. Beck (1968, 1973) suggested that the caves had developed preferentially within calcareous beds in the Glen Rose and that large caves, such as the Natural Bridge Caverns System and Bracken Bat Cave have developed parallel to the high-velocity flow of groundwater caused by a convergence of flowpaths in the vicinity of these caves. Beck based this interpretation on maps of the piezometric surface constructed by Rhodes and Guyton (1955). Features mapped and interpreted during the present study, and recent hypotheses on the evolution of the cavernous Edwards Aquifer, suggest a new explanation for the origin of the Natural Bridge Caverns System and nearby caves, as described below.

Lithologic Control

The concordance of passages with stratigraphic horizons in the Natural Bridge Caverns System, Bracken Bat Cave, and Double Decker Cave has been well documented and suggests strong lithologic control (Kastning, 1978, 1980e; Knox, 1979, 1981). The approximate, gradational contact between the Kainer Formation of the Edwards Group and the underlying Walnut Formation was determined from published described sections (see Appendix B for sources), from geologic maps by Newcomb (1971) and Abbott (1973), and from stratigraphic data taken within the caves. The Walnut and Glen Rose Formations, separated by an unconformity to the north, intercalate in southern Comal County (Moore, 1964, p. 6), and the contact between them is difficult to map, particularly inside caves. The Walnut-Glen Rose contact was chosen to lie along a brown, clay-rich bed, 1 cm thick, visible in the wall of the excavated tunnel just south of Pluto's Anteroom in Natural Bridge Caverns. This gives the Walnut Formation a thickness of 13.7 m in this vicinity.

Correlation of stratigraphic horizons within which major passages of the caves have developed is shown in Figure 119. Extensive passages of Bracken Bat Cave, the Natural Bridge Caverns System, and Double Decker Cave have developed in the Walnut Formation. In Natural Bridge Caverns and Double Decker Cave passages in the Walnut extend upward in places into the basal beds of the Edwards Limestone. This is the result of collapse of ceilings and upward stoping of passages (see below). These two caves also have large passages developed within the upper member of the Glen Rose Formation. The largest of the passages in the New Braunfels area is the one forming the lowest levels of the Natural Bridge Caverns System along its entire extent. It lies in the interval from 21 to 33 m below the contact between the Walnut and Glen Rose Formations (Figure 119).

Development of large passages in the Glen Rose and Walnut Formations may be attributed to enhancement of porosity following Early Cretaceous deposition, when the San Marcos Arch was elevated, and more than 30 m of the uppermost beds of the Edwards Group were removed by subaerial erosion (Rose, 1968,

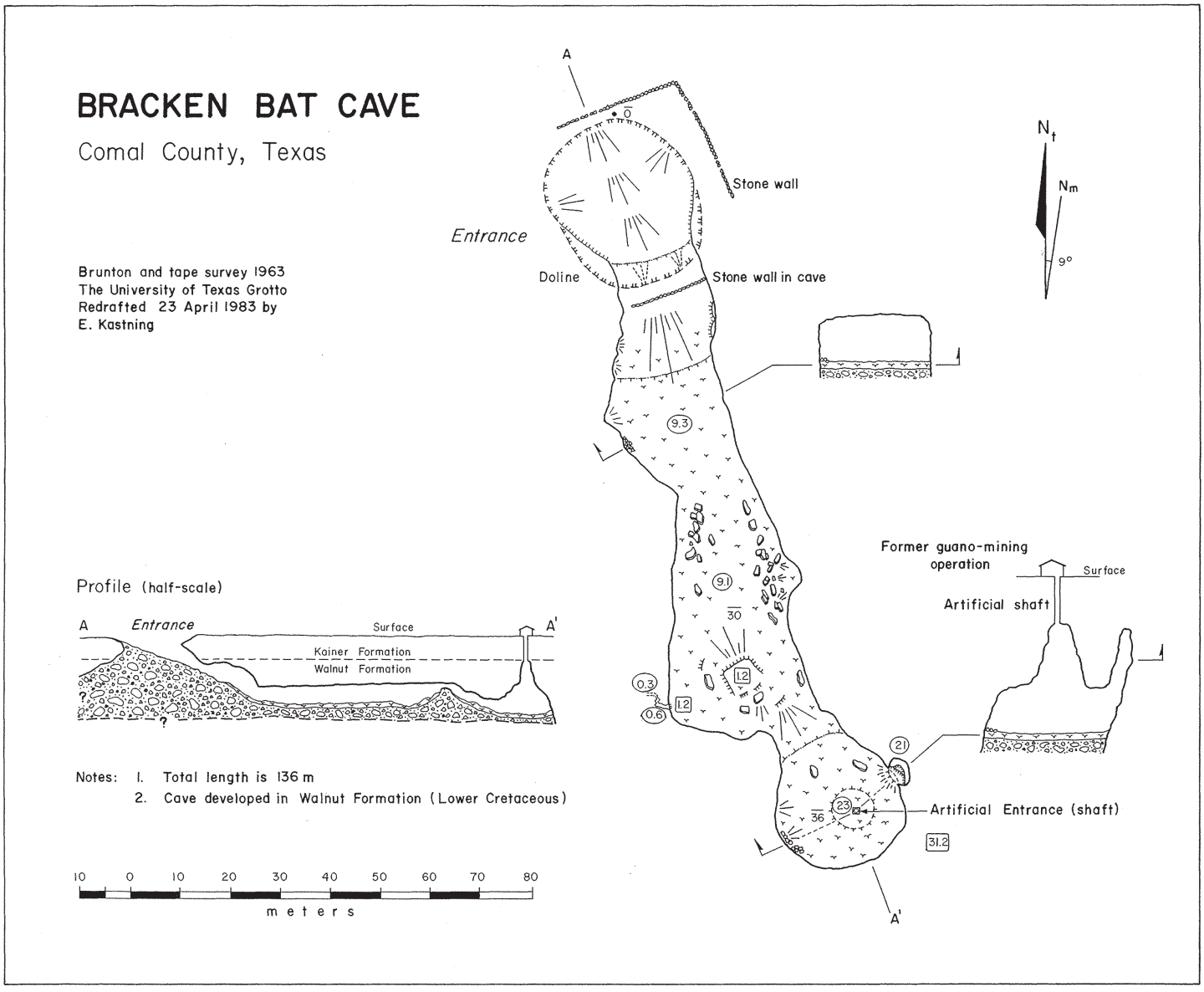


Figure 120: Map of Bracken Bat Cave. Modified from Reddell (1964a,b, p. 15) and Fieseler and others (1978, p. 111).

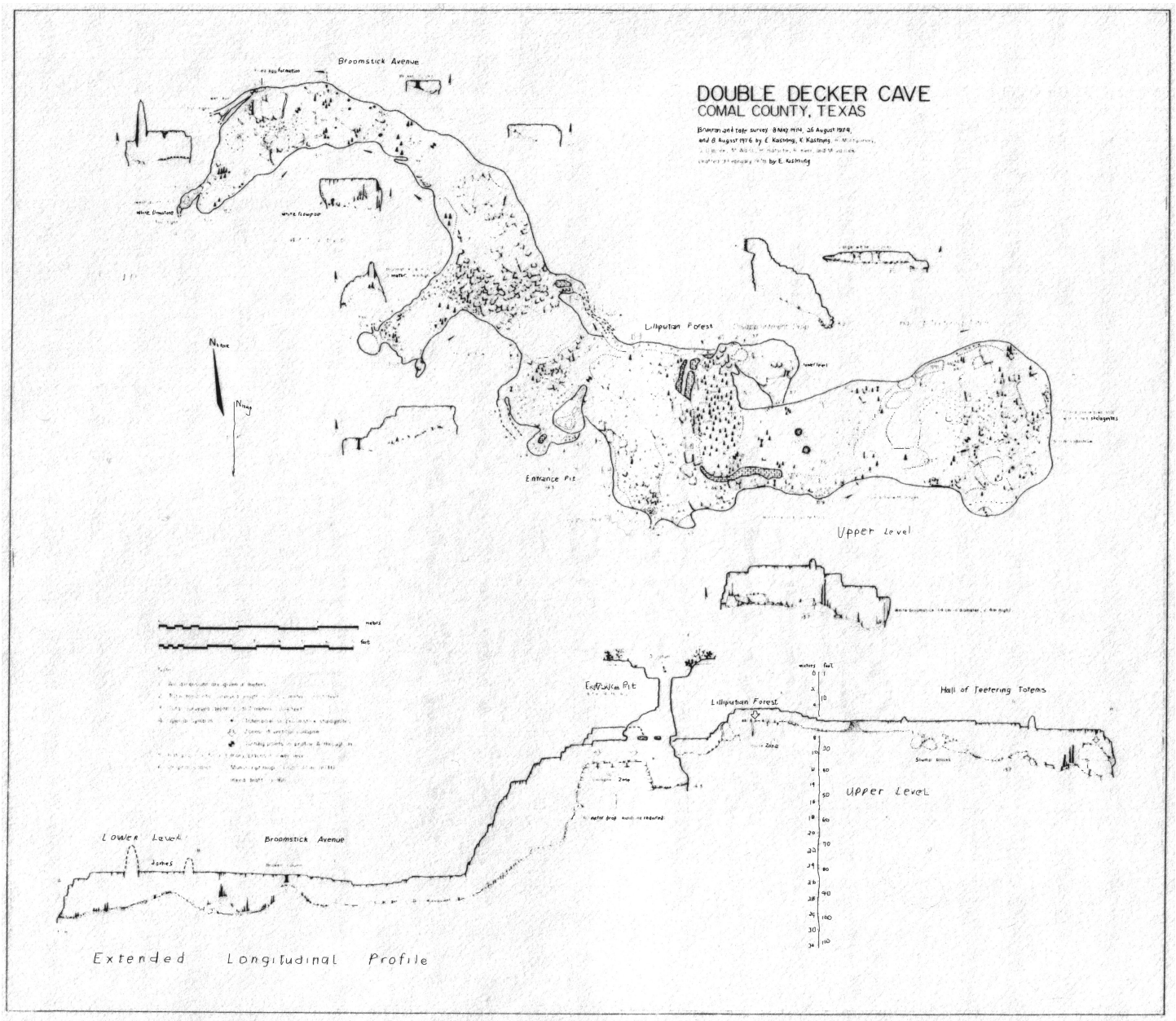


Figure 121: Map of Double Decker Cave.

1972). The accompanying increase in circulation of meteoric groundwater within the lower Edwards Group and underlying Walnut and Glen Rose Formation enlarged available pores.

Evidence for development of cavernous porosity predating development of integrated caves can be seen in Natural Bridge Caverns (Figures 122 and 123). Cavities in the upper member of the Glen Rose were enlarged during the early stages of circulation of groundwater and were subsequently filled with calcareous clay and marl. Laminations are easily seen in these deposits. The fill shown in Figure 122 is nearly as hard as the surrounding bedrock, and the solution-sculpted surface of the cave wall is continuous across the deposits within the cavity. Vugs lined with crystals of calcite occur in the Walnut Formation along the walls of the excavated tunnel leading from the entrance-chamber to St. Mary's Hall.

Some beds of the upper member of the Glen Rose Formation are highly porous in comparison with others above and below and are readily seen in walls of passages on most levels of the caves (Figures 124, 125, and 126). Much of this porosity has resulted from leaching of burrows, and by dissolution of evaporites and fossils (moldic porosity). Undoubtedly circulation of groundwater during the mid-Cretaceous enhanced much of this porosity, but most enlarged cavities were presumably small and poorly integrated in comparison to those that developed later.

The greatest effect of enhanced porosity from the mid-Cretaceous occurred during development of caves following uplift of the Edwards Plateau, Balcones faulting, and regional dissection. Groundwater was transmitted to porous zones through joints created by the faulting and along favorable bedding-plane partings. Zones of primary and enhanced porosity were subsequently integrated into a network of conduits. As a result, passages of caves of this area typically occupy stratigraphic horizons that originally had a high effective porosity.

Some stratigraphic control has also resulted from the mineralogic content of the host-rocks. Dolomitic beds are generally poor formers of caves in the presence of strata of greater calcitic content (Rauch and White, 1970; Rauch, 1972; Kastning, 1975). This is true of the caves of the New Braunfels area, where passages commonly occupy calcareous beds rather than dolomitic beds (Beck, 1968). However, the massive, dolomitic unit comprising the uppermost member of the Glen Rose does contain some passages. Moreover, selectivity of calcareous beds over dolomitic beds is prevalent only among strata of uniform porosity and permeability because these have been the dominant, lithologic determinants for the vertical positioning of passages. This is exemplified in Figure 125 where a dolomitic bed, sandwiched between two very porous beds, apparently has not inhibited development of the passage.

Structural Control

Each of the three caves studied in detail exhibit linearity in plan- and profile-view, suggesting strong, structural control (Figures 116, 120, and 121). Detailed mapping of the caves and their geologic structure has shown that fractures such as bedding-plane partings, joints, and faults have guided the orientation and

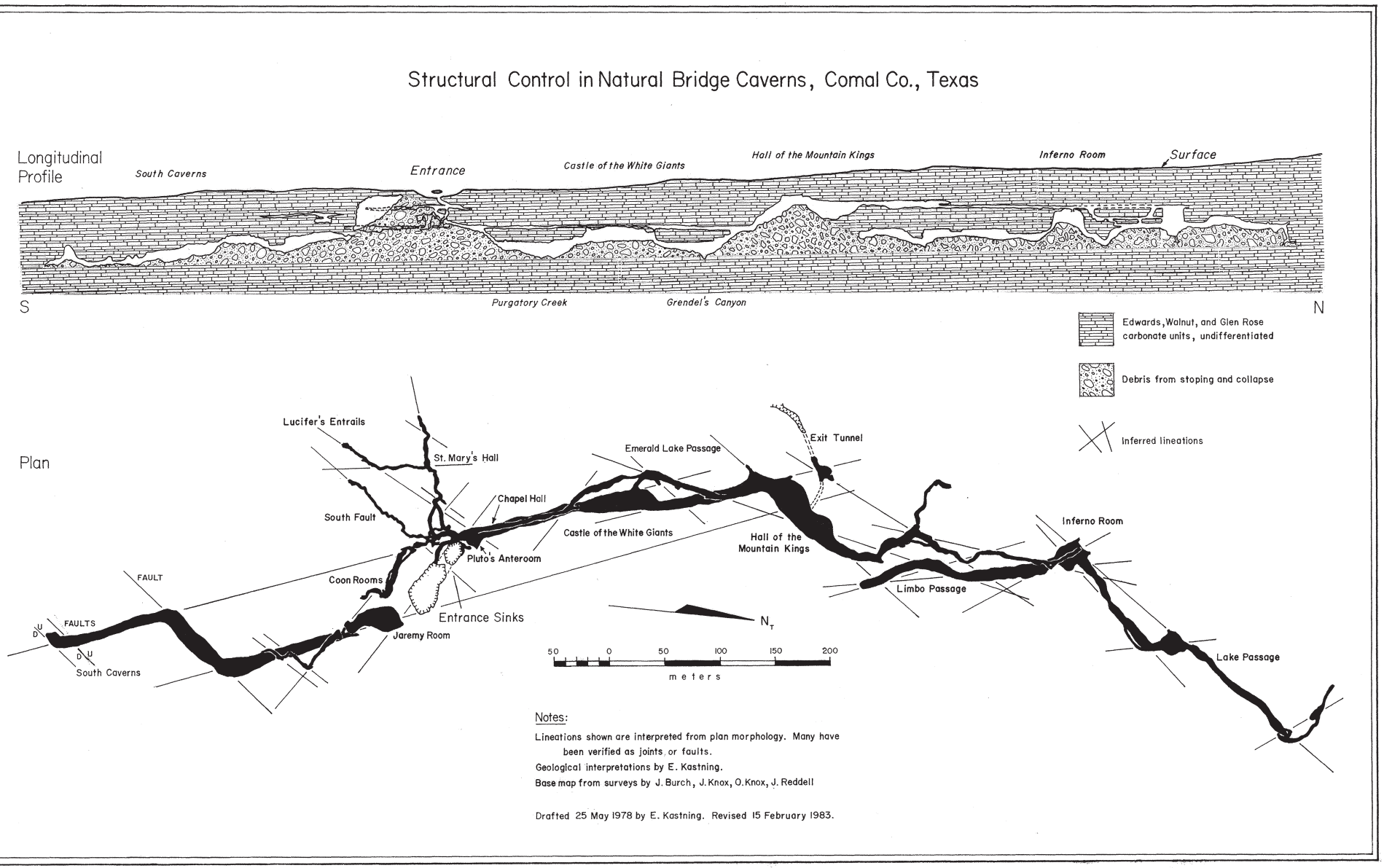


Figure 127: Structural control in Natural Bridge Caverns. Top figure shows zones of collapse. Bottom figure identifies lineations in plan-view, many of which correspond to mapped fractures. From Kastning (1978, p. 37).





Figure 122 (above)**:** Solutional cavity in the western wall of Pluto's Anteroom, Natural Bridge Caverns. Fill consists of hard marl. Cavity predates development of the cave.

Figure 123 (right)**:** Solutional cavity in the western wall of Emerald Lake Passage, Natural Bridge Caverns. Fill consists of soft clay and silt. Cavity predates development of this passage.



Figure 126: Chapel Hall in Natural Bridge Caverns. Passage has developed within porous beds and has been guided by the joint visible in the ceiling.

Figure 124 (below)**:** Exposure of porous limestone along the eastern wall of Castle of the White Giants, Natural Bridge Caverns. Pores resulted from leaching of burrows of pelycepods.

Figure 125 (right)**:** Highly porous beds at northern end of Emerald Lake Passage, Natural Bridge Caverns. A massive, dolomitic unit is interbedded between the porous strata. View is to the north.



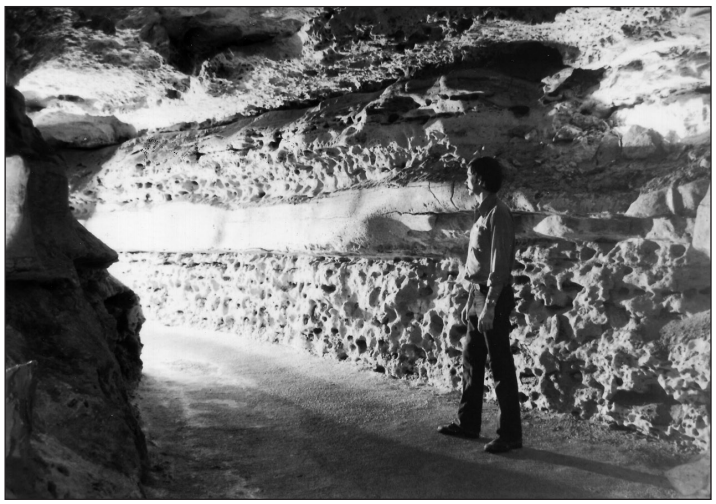




Figure 128: Dripstone formed by seepage along a bedding-plane parting, Pluto's Anteroom, Natural Bridge Caverns. View is to the north. Coin in top, center gives scale.

morphology of passages (Figure 127).

Bedding-Plane Partings. A few small passages have developed along bedding-plane partings with relatively little control by joints. Examples include meandering segments of the upper two levels in Natural Bridge Caverns (Coon Rooms, Jan's Long Crawl, and the Moors). The widths of passages commonly exceed their height, and Jan's Long Crawl is characterized by anastomosing crawlways averaging 0.3 to 0.4 m in height. The crawlway between Lilliputian Forest and Hall of Teetering Totems in Double Decker Cave may have originated along a bedding plane, but it has since been modified by collapse.

Horizontal permeability along bedding-plane partings is evident in the cave where movement of groundwater can be observed. Purgatory Creek and River Styx in Natural Bridge Caverns occupy phreatic tubes oriented along such partings, and some speleothems have formed where seepage enters the cave along partings (Figure 128). Some seepage entering along the walls of the cave does so along the top surfaces of shaly beds where water has been perched.

Joints. Joints have exerted the greatest structural control on the morphology of passages. Joints are readily identified in ceilings of passages in all three caves. Most linear segments of passages in the Natural Bridge Caverns System have been verified as joints (Figures 126 and 127).

Joints measured in the Natural Bridge Caverns System have two major trends: N 20°-30° W and N 40°-50° E (Figure 127). The northeastern joint-set is parallel to the Bat Cave Fault (Figure 115) and is a product of Balcones faulting. The northwestern set is not compatible with tensional stresses operating during normal (step) faulting. However, the strike of this set is nearly parallel to the axis of the San Marcos Arch, and joints of this orientation appear to be related to uplift along the San Marcos Platform during the Cretaceous. The northwestern joint-set accounts for 35 percent of the total length of passages in the Natural Bridge Caverns System, whereas the northeastern set represents 15 percent of this length. The remaining 50 percent of the length has been controlled by shorter joints of other orientations, by faults, or by bedding-plane partings.

Trends of most joints observed in Bracken Bat Cave lie in the intervals N 10°-30° W and N 0°-10° E. Joints of the former set trend parallel to the primary northwestern joint-set of the Natural Bridge Caverns System. Joints of the northeastern set, although present, account for little length of passages in this cave. The lower level of Double Decker Cave is mostly oriented along the dominant, northwestern set of joints. The upper level, however, is oriented along joints trending between N 80° E and S 70° E. The trend of this set is parallel to the Zuercher Ranch Fault (strike of N 85° E) located just 275 m north of the cave (Figure 115).

Control by Faults. High-angle faults have been observed within the South Caverns of the Natural Bridge Caverns System and in Double Decker Cave (Figures 129 and 130). Faults in the Natural Bridge Caverns System are parallel to the Bat Cave Fault, and their throws generally measure less than a meter (Figure 129). A single fault in Double Decker Cave strikes parallel to the Zuercher Ranch Fault, and its throw is approximately 0.5 m.

Faults can enhance or inhibit the circulation of groundwater in carbonate aquifers, depending on (1) the type of faulting: normal versus thrust, (2) the type of stresses operating during faulting: tensional versus compressional, (3) the lithologies in juxtaposition across faulted strata, (4) the density of joints related to faults, and (5) the degree of brecciation or shatter and the thickness of such a zone (Kastning, 1977c). Faults may also be of neutral influence. Thus, no general statement on the influence of faults on the flow of groundwater or on speleogenesis is universally valid (Kastning, 1977c). Oddly enough, examples of positive, neutral, and negative effects of faults on development of passages occur in the Natural Bridge Caverns System and Double Decker Cave. The southernmost 200 m of the South Caverns alone has been affected in each of these ways (Figure 131).

Box Canyon in the South Caverns has developed along a normal fault with a throw of 0.6 m. The fault is clearly visible in the wall at the southwestern end of this segment of passage where the cave turns to the southeast. Joints of narrow spacing, visible in the ceiling of Box Canyon, suggest that tensional stress during faulting created a zone of enhanced permeability which became integrated with joints of the prominent northwestern set. The master conduit then developed along this integrated flowpath. The fact that the width of Box Canyon is greater than that of other segments of passages connecting with it may be explained by the high density of joints of this fault.

In the Fault Room to the south two visible faults transect the cave, but have no expression in the morphology of the passage (Figure 129). Apparently the small number of associated joints precluded enhancement of permeability, yet the movement of groundwater across the faults remained uninterrupted because displacements were small (0.3 to 0.6 m) and lithologies on both sides of the faults remained similar

The main conduit of the Natural Bridge Caverns System ends abruptly at the southern terminus of the South Caverns. Comparison of the map of the cave (Figure 116) with the geologic map (Figure 115) shows that this termination is coincident

Figure 129: *En echelon* faults in Fault Room of South Caverns, Natural Bridge Caverns System. Passage terminates approximately 20 m to the right. View is to the east.



with the Bat Cave Fault. The throw of this fault is approximately 70 m in this vicinity (Figure 131), and beds of the Kainer Formation are in juxtaposition with the upper member of the Glen Rose Formation. The lithologic change across the fault prevented the flow of groundwater across it and to the south. Presumably flow at this location became diverted to the northeast, in conjunction with the regional movement of groundwater to points of discharge at major springs along the Balcones Escarpment. The Bat Cave Fault is the southeastern boundary of the compartment in which groundwater of the area of the Natural Bridge Caverns System moves. There is no visible conduit leading from the Fault Room, apparently because it lies below the floor of breakdown and sediment.

Flowstone is being deposited where the fault transects Double Decker Cave at the drop between the Upper and Lower Levels (Figure 130). Here water from the surface infiltrates into the bedrock along the fault-plane and enters the cave as seepage.

Correlation of Caves with Surficial Fractures. Fractures were mapped from remotely sensed imagery of the New Braunfels area in order to ascertain how well joints and faults can be interpreted from such imagery, and to determine whether this method is helpful in analyzing karstic landforms and speleogenesis. A 6.89- by 9.93-km rectangular area (Figures 115 and 132) was chosen because (1) a wide variety of remotely sensed imagery was available for this location, (2) the area overlies the Balcones Fault Zone where geologic and hydrologic conditions are diverse, (3) the area is part of the recharge-zone of the Edwards Aquifer, (4) the geology of the quadrangle has been mapped in detail (George and others, 1947. 1952; Newcomb, 1971; Barnes, 1974b), (5) significant studies of groundwater have been published

Figure 130: Flowstone on southwestern wall of Double Decker Cave, approximately 12 m northwest of the entrance. Seepage enters the cave along the fault visible at the top of the flowstone. This speleothem is approximately 2.5 m long.



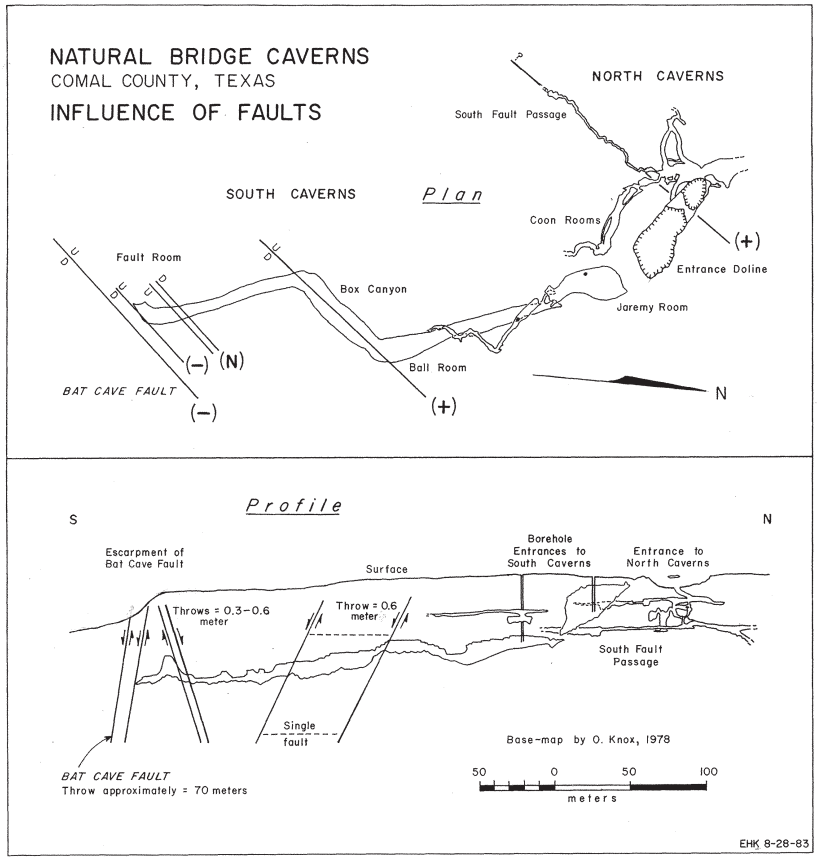


Figure 131: Plan and profile of the South Caverns, Natural Bridge Caverns System, showing faults. Plus sign ( ) indicates fault has provided permeable zone for speleogenesis. Negative sign (-) indicates fault has prevented flow of groundwater across it. N indicates fault has not affected flow of groundwater during speleogenesis. Throws are indicated in meters. Modified from map of Figure 116.

on the area (Livingston and others, 1936; George and others, 1947, 1952; Green, 1967; Abbott, 1973, 1977a,b), (6) caves of the area have been precisely surveyed (Reddell, 1964a,b); Kastning, 1978), and (7) structural elements have been mapped in three large caves (Kastning, 1978, 1981b, 1984).

Available imagery included: (1) low-altitude, black-and-white (panchromatic) prints, (2) a controlled photomosaic, (3) high-altitude, color-infrared transparencies, and (4) black-and-white (Band 5) LANDSAT (ERTS) imagery. Types of imagery, scale, date, source, and roll and frame information are summarized in Appendix E. Resolution of geologic features on the imagery varied widely depending on photographic media (satellite, panchromatic, or color infrared), altitude of flight, and scale used. Appendix E evaluates the four types of imagery and their suitability for identifying and mapping photolineaments which presumably represent traces of fractures.

Lineaments mapped from six sets of imagery (four types) have been plotted on a single map (Figure 132). As indicated by the map, resolution of lineaments is variable from one set of imagery to another. The reasons for this are discussed in Appendix E.

Lineaments are visible where fractured rocks crop out or where they are near enough to the surface to transmit structural patters through the overburden. Lineaments commonly reflect (1) changes in substrate, regolith, and soil in the vicinity of fractures, (2) patterns in vegetation due to availability of moisture in fractures, (3) difference sin moisture within surficial material on opposite sides of a fracture, and (4) changes in the albedo of the surface owing to retention of moisture in fractures. Although most fractures can be discerned on the ground, some are too extensive or subtle to be recognized easily in this way. However, these may be readily detected and mapped from aerial, suborbital, or orbital platforms.

A comparison of the map of fractures (Figure 132) with the geologic map (Figure 115) shows that many lineaments parallel the Hidden Valley, Zaccaria Ranch, Zuercher Ranch, Bat Cave and Hueco Springs Faults of the Balcones Fault Zone. A second prominent set of lineaments trending N 10°-40° W represents fractures associated with the San Marcos Arch. Numerous shorter lineaments on the map are orthogonal in fractures of the two prominent sets, and may be regarded as conjugate

fractures associated with the Balcones Fault Zone and San Marcos Arch.

Orientations of mapped photolineaments (Figure 132) and of linear segments of passage in caves (Figure 115, 120, 121, 127, and 131) may be compared form rose-diagram (Figure 133). It is readily seen that the orientations of passages have been largely guided by fractures.

Modification by Collapse. Each of the three caves discussed in this chapter exhibits significant modification through the collapse of ceilings of the original conduits. Nearly all of the large chambers of the Natural Bridge Caverns System, including the Fault Room, Ball Room, Jaremy Room, Pluto's Anteroom, Castle of the White Giants, Hall of the Mountain Kings, Inferno Room, and Dome Pit, are floored by mounds of debris from collapse (Figures 116, 127, and 134). Collapsed blocks are typical of the types of breakdown observed in caves (Davies, 1949). Zones of collapse are evident in the vicinity of the Entrance Pit and along the Upper Level of Double Decker Cave (Figure 121). The slope within the doline at the entrance to Bracken Bat Cave is floored with debris from collapse, and a mound of breakdown can be found 80 m into the cave, just before the constriction in the passage leading to the final room (Figure 120).

Presumably, buoyant facies provided by phreatic water were removed as the caves drained, allowing collapse of ceiling previously weakened by joints and solutional enlargement of conduits. Continued collapse promoted upward sloping of passages until mechanical stresses were stabilized in accordance with the present configurations of passages. Stoping occurred along the entire length of the master conduit in the lower levels of the Natural Bridge Caverns System (Figure 127), but large room of collapse, termed "breakout-domes", are localized in areas of intense fracturing. Commonly such rooms have formed where two or more prominent fractures intersect (Figure 127)., suggesting that such intersections greatly weaken beds of the ceiling. Davies (1951) has presented a mathematical and mechanical analysis of the formation of breakout domes. The Hall of the Mountain Kings (Figure 134) is a "textbook example" of a breakout-dome, similar in form to those found in Mammoth Cave in Kentucky (White and White, 1969) and Wynadotte Cave of Indiana (Malott, 1951, p. 33). Stoping in Natural Bride Caverns has transgressed beds of the Walnut Formation and progressed 6 m into the overlying Kainer Formation.

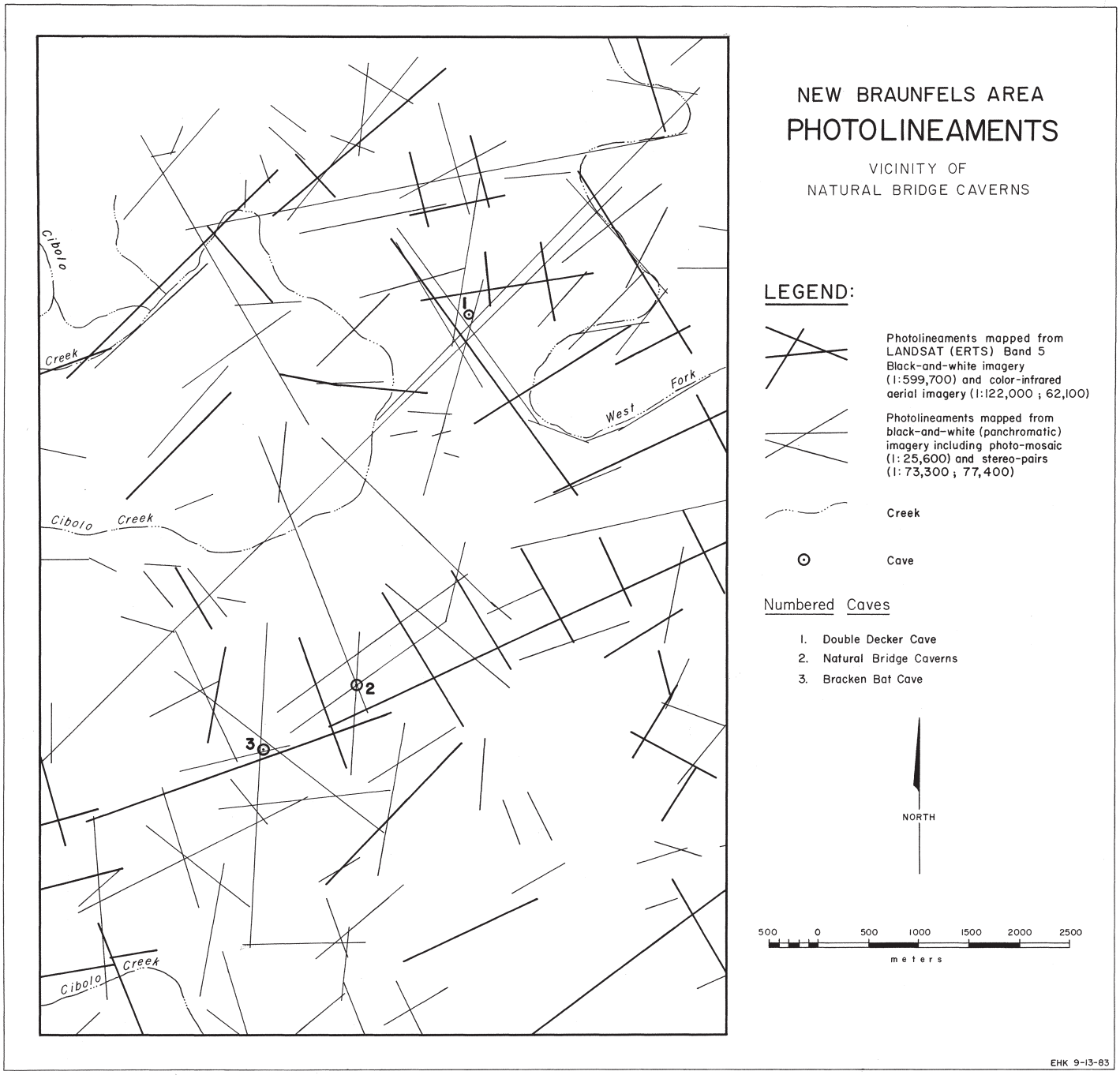


Figure 132: Fractures mapped from remotely sensed imagery, New Braunfels area. Region shown is that of Figure 115. Heavy lines indicate fractures mapped from LANDSAT and color, infrared imagery. Light lines indicate fractures mapped from black-and-white, panchromatic imagery. See Appendix E for evaluation of various forms of imagery used in mapping fractures.

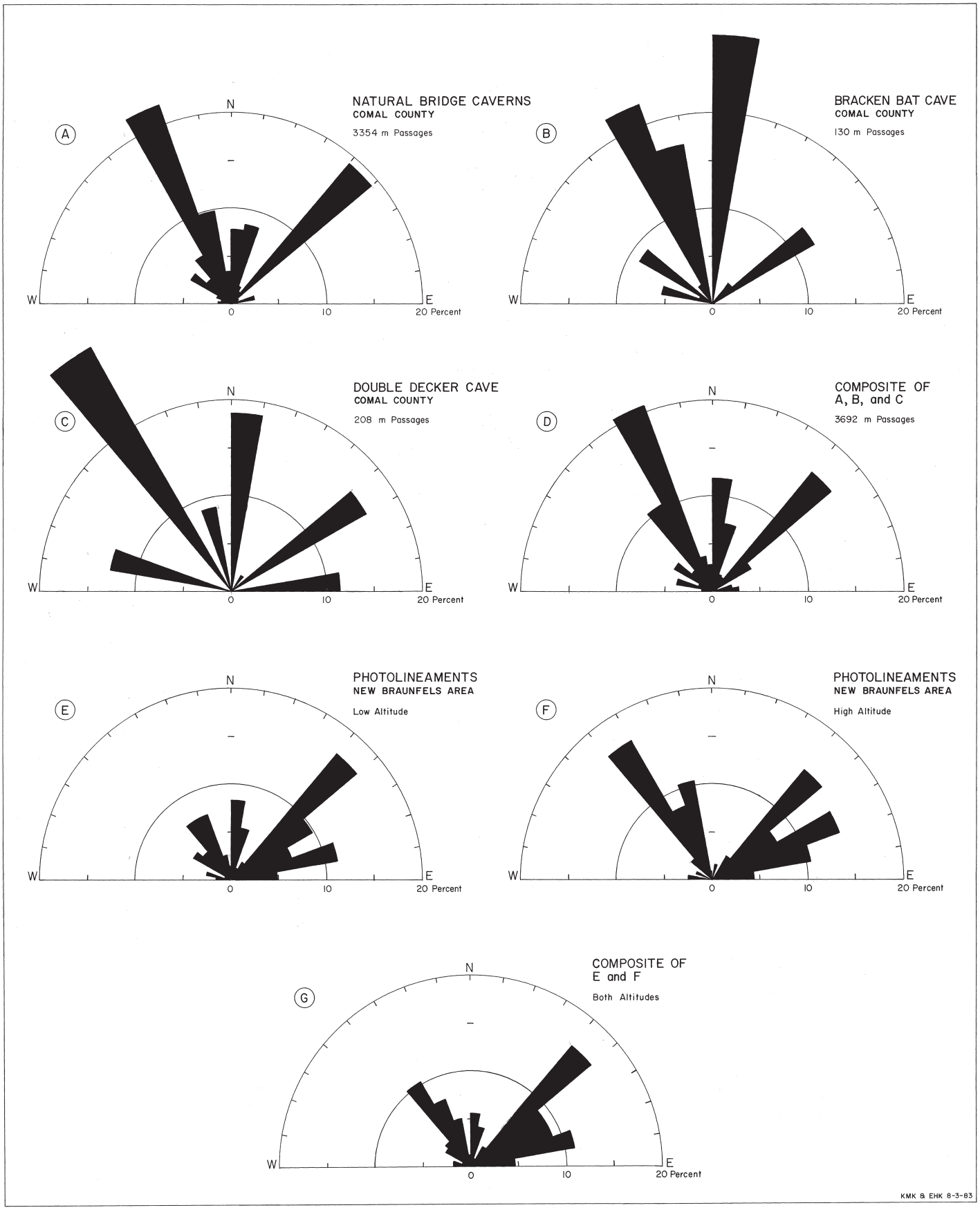


Figure 133: Rose-diagrams of linear segments of passages in (A) Natural Bridge Caverns System, (B) Bracken Bat Cave, (C) Double Decker Cave, and (D) all three caves; and of photolineaments mapped from aerial photographs flown at (E) low altitude, (F) high altitude, and (G) both altitudes. *See also* Figures 116, 119, 120, and 132.

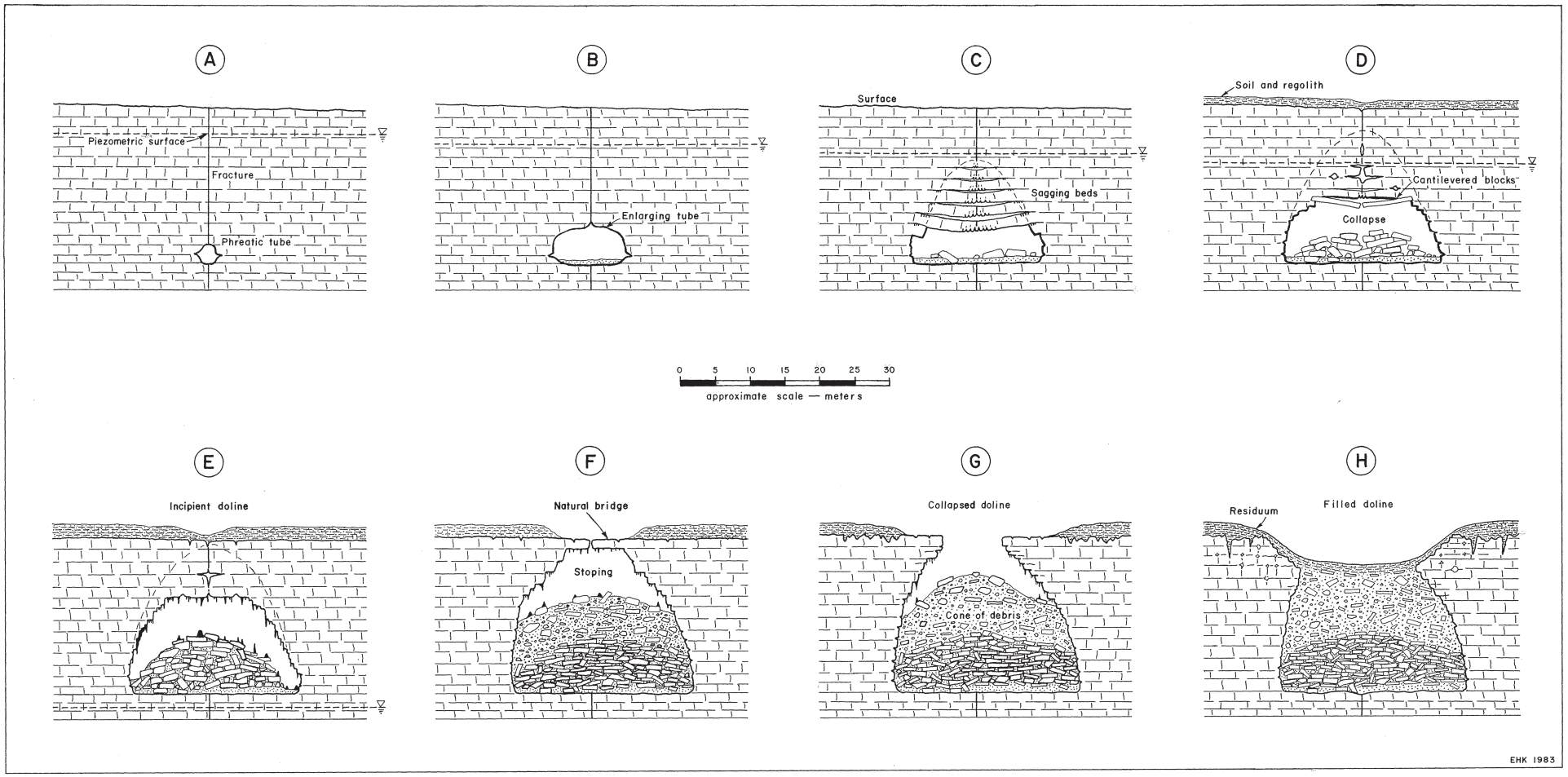


Figure 135: Schematic diagram illustrating the evolution of breakout-domes in the Natural Bridge Caverns System.

Ceilings are remarkably flat and elliptical, and walls are flared in a mechanically stable configuration.

The most massive collapse and stoping occurred in the vicinity of the entrance to Natural Bridge Caverns. Here stoping progressed from the Glen Rose Formation, through the overlying Walnut and Kainer Formations, and intersected the surface, forming the doline at the entrance. The Natural Bridge, namesake of the cave, is a remnant which has not yet collapsed (Figure 118). The northern end of the South Caverns, Jaremy Room, Pluto's Anteroom, and Sherwood Forest all lie on the flanks of the large mound of debris remaining from collapse (Figures 127). The doline at the entrance to Bracken Bat Cave has formed in an identical manner.

The volume of debris in mounds within the breakout-domes is generally less than that of the material which has collapsed. This suggests that much of the debris has been removed by dissolution. Waters may have gradually drained from the caves over a considerably period of time so that, as buoyant forces were removed, there was still sufficient circulation of water through the lower parts of the chambers to remove material from collapse. Alternatively, vertical fluctuations of waterlevels may have been sufficient over a long time to allow collapse and concomitant dissolutional removal of debris. The genetic sequence for the development of breakout-domes in caves of the New Braunfels area is schematically shown in Figure 135.

Collapse has greatly modified the flow of vadose water through conduits of Natural Bridge Caverns. River Styx and Purgatory Creek have been diverted away from the master conduit in areas of collapse, and emerge only in the lowest points of the cave where collapse has been minimal. Water rising form these streams during severe flooding is retained behind dams created by the mounds of debris.

Speleothems

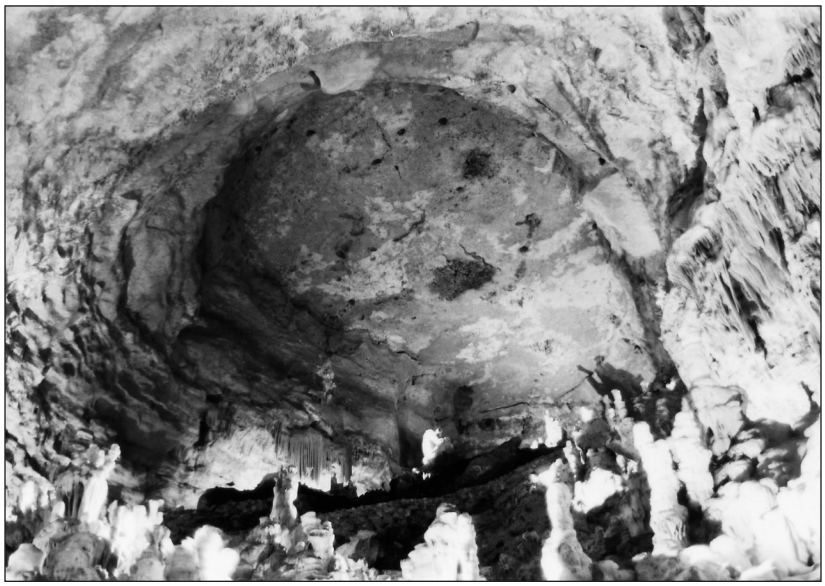
Speleothems in the Natural Bridge Caverns System range in size from massive stalagmites and mounds of flowstone (Figure 117) to slender, sodastraw-stalactites. Sodastraws in the Fault Room average 0.5 to 2 m in length and one specimen, suspended form the high ceiling of this room, is over 4.3 m in length, one of the longest known in the world. Sodastraws form comparatively rapidly (as much as a few millimeters in length per year, and these formations are relatively recent features associated with slow, vertical seepage along joints. The massive formations of the Castle of the White Giants, on the other hands, are substantially older and still actively forming.

As with other speleothems in caves of the Edwards Plateau many of those which are no longer active show evidence of redissolution. Redissolved stalagmites are common at all levels of the cave, but are more numerous in the Hall of the Mountain Kings (Figure 136). Some stalagmites have been reshaped into streamlined forms by flowing water. The large stalagmites in the Castle of the White Giants have solutional hollows created by redissolutions (Figure 117).

Redissolved formations are found as high in the cave as the top of the Walnut Formation in the Hall of the Mountain Kings (Figure 136). This indicates that the cave flooded to this level after it was originally drained of phreatic water, collapse had occurred, and enough time had elapsed for deposition of large speleothems. The most plausible mechanism for such redissolution is the introduction of colder and wetter climate in the area, thus increasing recharge to the cave system. This condition probably coincided with glacial stages of the Pleistocene when central Texas was subjected to a pluvial climate.

Flooding during pluvial periods was probably episodic, corresponding to periods of high precipitation. This is suggested by small, incised channels at the bases of solutional hollows in formations in the Castle of the White Giants, giving each niche the shape of a keyhole (Figure 117). The channels formed as water slowly receded following each flood and drained from these hollows

Figure 134: Breakout-dome in Hall of the Mountain Kings, Natural Bridge Caverns. Stonewalls have been constructed along a trail winding up the mound of debris in the foreground. Uppermost beds visible in the dome are of the Edwards Group. The remainder in this view are of the Walnut Formation. View is to the northeast.



Chronology of Speleogenesis

The developmental history of the three caves proceeded according to the following chronological sequence:

1. The San Marcos Platform was raised episodically along its northwestern axis throughout the post-Albian Cretaceous, producing a system of fractures parallel to its axis. Subaerial exposure of the platform in the late Albian resulted in removal of over 30 m of uppermost beds of the Edwards Group, steepening of hydraulic gradients, and circulation of groundwater through fractures. Flow of groundwater was primarily to the south, and the original pores were subsequently enlarged into poorly integrated cavities (Woodruff and Abbott, 1979a,b).

2. During the remainder of the Cretaceous, the region was alternately covered by shallow, marine, shelfal waters and exposed to subaerial conditions. Indications of the latter include (1) absence of the Del Rio Clay just west of New Braunfels (where the Buda Limestone rests unconformably on the Georgetown Formation), (2) absence of the lower members of the Austin Chalk on the San Marcos Arch, and (3) presence of Upper Cretaceous rocks (e.g. Austin Chalk) as fill within dolines in units older than those of the Eagle Ford Group (Dumble, 1918, p. 19-20; Adkins, 1932, p. 361, 401; George and others, 1947, p. 33; 1952, p. 34). Solutional enlargement lessened or ceased during periods of submergence as many zones of permeability were previously sealed in the Early Cretaceous with argillaceous and micritic sediments. Many cavities were filled with marl, micrite, and detritus. Although further accentuation of the San Marcos Arch in the Late Cretaceous continued to produce fractures, speleogenesis in the area of the Natural Bridge Caverns System was substantially curtailed under conditions of reduced porosity and in the absence of steep hydraulic gradients in the evolving aquifer.

3. The region was lifted above sealevel near the close of the Cretaceous, and rocks attained a gentle (less than 0.5°) southeastward dip. However, an absence of nearby points of discharge precluded development of sizeable solutional conduits.

4. Regional (Balcones) faulting during the Miocene significantly accelerated the geomorphic evolution of drainage on the surface and in the subsurface. Northeast-trending fractures produced during faulting were overprinted on the pre-existing northwest-oriented set of fractures. The Balcones Escarpment established a low baselevel to the southeast toward which surficial and subsurficial geomorphic processes began to operate.

5. Flow of groundwater increased in response to steepened hydraulic gradients created by drainage to springs at baselevel along the Balcones Escarpment. Some faults, such as the Bat Cave Fault, imposed impermeable boundaries on the aquifer, diverting groundwater along the strike of the faults. Flow along fractures and bedding-plane partings within the phreatic zone became integrated into master conduits, producing some of the larger passages of caves. These include the main passage along the lower level of the Natural Bridge Caverns System, the large passage of Bracken Bat Cave, and the Upper and Lower Levels of Double Decker Cave. Some conduits followed beds of favorable primary and solutionally enhanced porosity, and became oriented down the dip, in the direction of greatest hydraulic gradient. However, other passages cut across previous solutional cavities that are now filled with pseudospar (K. Young, personal communication). Most conduits were strongly guided by fractures of the two primary sets. Formed by tensional stresses during uplift of the San Marcos Arch and faulting along the Balcones Fault Zone, these fractures were sufficiently open to easily accommodate phreatic flow. In places, dip-oriented conduits at different levels were integrated through vertical chimneys along fractures, allowing flow to move continuously through a single solutional channel from one stratigraphic horizon to the next. This situation is exemplified by Double Decker Cave where flow was communicated from one level



Figure 136: Stereoscopic photograph of redissolved formations near the exit-tunnel atop the mound of debris, Hall of the Mountain Kings, Natural Bridge Caverns.

to the other along a major fault (Figure 130).

6. As conduits enlarged, their widths became too great to support overlying beds, and collapse of ceilings began. This may have been initiated as water first began to drain from the caves, removing bouyant support of the ceilings.

7. The piezometric surface in the aquifer began to drop as surficial streams became entrenched along the escarpment and as a consequence of the capture of streams (Woodruff and Abbott, 1979a,b). The caves began to drain as the dropping, piezometric surface intersected the conduits. Mounds of debris in the caves were partially removed by vadose waters still flowing in the conduits. Areas of collapse became mechanically stabilized.

8. Speleothem deposition began.

9. Increased precipitation and decreased evaporation during pluvial climates of the Pleistocene resulted in high discharges in the existing caves. Collapse and alluviation within the caves prevented efficient throughflow of groundwater. As a consequence, caves underwent periodic flooding which redissolved speleothems and locally promoted development of routes of bypass for floodwater around mounds of debris previously created by collapse. Smaller, middle- and upper-level passages in the Natural Bridge Caverns System developed as routes of bypass in response to blockages at the Fault Room, the area between Pluto's Anteroom and Sherwood Forest, the southern end of Grendel's Canyon, the southern end of the Limbo Passage, and the Inferno Room. Pre-existing conduits were also enlarged by floodwaters.

10. Vadose flow in the caves diminished to present levels with the advent of warmer climates. This flow apparently represents underflow derived from recharge along Cibolo Creek. Under conditions of baseflow, discharge is carried through inaccessible conduits beneath the explored levels of Natural Bridge Caverns.

SUMMARY

The development of caves in the New Braunfels area progressed in response to two distinct tectonic episodes. First, uplift of the San Marcos Arch subjected Lower Cretaceous rocks to subaerial exposure and heightened topographic relief that promoted deep circulation of groundwater through fractures produced during the tectonism. Primary porosity in the Glen Rose Formation became enhanced as cavities were solutionally enlarged; but openings remained poorly integrated. During the Late Cretaceous, the region was episodically covered by shallow seas which deposited calcareous, siliclastic, and marly sediments above the Glen Rose Formation. The second disturbance, Balcones faulting during the Miocene, heightened local relief in the immediate area and thus steepened hydraulic gradients. Groundwater moving along the gentle southeastern dip enlarged pre-existing, northwest-trending fractures as well as many of those produced by the faulting. As a result, major caves of the area, such as the Natural Bridge Caverns System, consist of well-integrated conduits of large cross-section that are angulate in plan-view and correspond to flowpaths favorably aligned along open fractures.

Incision of streams draining the Edwards Plateau was rapid in the vicinity of the Balcones Escarpment. Levels of groundwater declined in response to cutting of valleys, and levels of passages in caves developed in a descending sequence. Passages were enlarged to canyon-like cross-section by subsequent vadose streams and later abandoned as water circulated at greater depths. Collapse of ceilings in some places blocked major, active conduits and promoted development of routes of diversion.

In general, groundwater flowed to the southeast. Moreover, northwest-trending fractures account for a large percentage of the overall orientation of cave passages. However, many northeast-trending fractures and some fractures of conjugate sets provided cross-overs in the paths of flow between adjacent, master fractures and thereby account for a significant number of passages as well.

11
AUSTIN—SAN MARCOS AREA

DESCRIPTION OF AREA

The Austin-San Marcos area lies largely within Travis, Hays, and Comal Counties (Plate 1). It is bounded on the south and north by lines of 29°50'00"and 30°22'30" north latitude respectively and on the east and west by lines of 97°42'30" and 98°10'00" west longitude respectively. The northern part of the area is drained by the Colorado River and the southern part by the Blanco and Guadalupe Rivers (Plate 2).

The geology of the area has been mapped at various scales ranging from 1:24,000 to 1:250,000, including regional maps (Barnes, 1974a,b,c; DeCook, 1960, 1963; Garner and Young, 1976; George and others, 1947, 1952; Hill and Vaughan, 1902; Trippet and Garner, 1977; Whitney and Young, 1957-1959; Woodruff, 1975) and maps of specific quadrangles and local areas (Bills, 1957; Bryant, 1948; Davis, 1962; DeCook, 1956; Grimshaw, 1970, 1976; King, 1957; Marquez, 1947; Nicholson, 1947; Noyes, 1957; Noyes and Young, 1960; Rodda and others, 1970). The structural setting of the Balcones Fault Zone has been investigated in detail in the Austin area (Balke, 1958; Caran and others, 1981; Dunaway, 1962; Kurie, 1956; McReynolds, 1958; Muehlberger and Kurie, 1956; Sandifer, 1935; Strong, 1957; Watkins and Muehlberger, 1962).

Like the New Braunfels area to the south (Chapter 10) the Austin-San Marcos area overlies the faulted outcrop of the Edwards Group (Plate 3). The outcrop of the Edwards, which narrows in the vicinity of Austin, is broken by numerous faults that are generally of shorter length and less throw than are faults to the south of San Marcos. However, the Mt. Bonnell Fault at Mount Bonnell in Austin has a stratigraphic throw of 220 m and the Edwards is faulted out completely (Sandifer, 1935; Balke, 1958; Rodda and others, 1970; Garner and Young, 1976). Upper Cretaceous units are exposed to the east of the Edwards outcrop and are commonly in juxtaposition with the latter along faults. Dissection along the Balcones Escarpment has exposed the Glen Rose Formation west of the fault zone. Ridges and hills are locally capped by the Walnut Formation and Edwards limestone. Individual units are described below in context with the geologic settings of particular caves.

Carbonate aquifers of this area are compartmentalized by step-faulting. In the southern part of the area near San Marcos and south of the Blanco River, much of the groundwater flows within the San Marcos Springs component of the Edwards Aquifer (Maclay and Small, 1976, 1983; Guyton and Associates, 1979). West of San Marcos and north of the Blanco River, groundwater is confined to smaller faulted blocks and emerges at springs with lower discharges than those to the south. These include Jacobs Well (discussed below) and Barton Springs. Few systematic studies have been completed on the character of the Edwards Aquifer north of San Marcos. Published information to date is limited to inventories of wells and water-levels (Follett, 1956e), quality of groundwater (St. Clair, 1979), and suitability of land for development (Woodruff, 1979).

Extensive faulting and fracturing of the Edwards in the area of outcrop has substantially increased the permeability of this formation. As a result streams crossing the outcrop commonly lose water through rapid infiltration. An excellent example is Barton Creek in Austin (Baker and Watson, 1974; Twidwell, 1976). The piezometric surface has declined in the upstream part of the losing reach, allowing an increase in available storage-space within fractures and solutional openings. Downstream in the losing reach the channel of Barton Creek intersects the piezometric surface, and the creek is fed by small springs and seeps during wet periods. Water issues from the Edwards at Barton Springs near the confluence of Barton Creek with the Colorado River. Here water discharges at 0.5 to 2.7 m^3/sec from solutional openings aligned along a fault (Brune, 1975). The area of catchment for Barton Springs has not been clearly defined; however, it is likely that recharge along the losing reach of Barton Creek and northeastward flow within faulted blocks of the Edwards Group contribute to the flow of the spring (Twidwell, 1976)

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

The Austin-San Marcos area contains many caves, yet few exceed 30 m in length. Caves are distributed throughout the area, but most, including the longest caves, are within the Balcones Fault Zone.

Most known caves of Travis and Comal Counties have been described by Reddell and Russell (1961a) and Reddell (1964a) respectively. Information on caves of Hays County is on file with the Texas Speleological Survey. Some of the larger and more significant caves have been described in the speleological literature, including Airman's Cave (Russell, 1975b), Cave X (Reddell and Russell, 1961a; Russell, 1974), Dead Dog Cave No. 2 (Fieseler, 1976), Wonder Cave (Elliott, 1970a; Russell, 1976), Ezell's Cave (Davis, 1971; Russell, 1976), Wimberly Bat Cave (Russell, 1973), and Jacobs Well (Bordelon, 1975; Jasek and Peck, 1975; Grimshaw, 1970). These are discussed below.

Frank (1964, 1965a, 1971) analyzed the sediments and vertebrate fossils from Fyllan Cave. Vertebrate fossils indicate that the age of deposits in the cave is Kansan (Patton, 1965). Collections have also been made in Cave X (Frank, 1961b, 1964).

Local, geologic studies have reported on karstic features, such as dolines, karren, and springs, in the Austin-San Marcos area (Davis, 1962; DeCook, 1956; Grimshaw, 1970, 1976; King, 1957; Noyes, 1957). Abbott (1973, 1975, 1977a,b) investigated the hydrogeology of the Edwards Aquifer in Comal and Hays Counties and identified an initial period of speleogenesis resulting from movement of groundwater and dissolution of limestone induced by uplift of the San Marcos Platform during the Cretaceous. The cavernous porosity resulting from this event was later enhanced and integrated into conduits during the Tertiary and Quaternary, in response to uplift of the Edwards Plateau in the Miocene. The hydrogeologic interrelationships among some of the caves in the San Marcos area were investigated in conjunction with the distribution of troglobytic faunas within caves, springs, and artificial wells intersecting solutional conduits (Uhlenhuth, 1921, 1948; Russell, 1976).

DEVELOPMENT OF CAVES

The emphasis of investigation in the Austin-San Marcos area was to analyze caves wherein passages appear to be morphologically typical of those found in the area, and are noteworthy because they reflect specific, geologic and hydrologic controls. Nine caves were selected, ranging in length from 13 m to 3.41 km. Five of these, Airman's Cave, Cave X, Bee Creek Cave, Dead Dog Cave No. 2, and Swimmin' Hole, are in the Austin area. Two, Wonder and Ezell's Caves, are in the San Marcos area, and two others, Wimberley Bat Cave and Jacobs Well, are near Wimberley.

Descriptions of Caves

Airman's Cave

Airman's Cave lies along Barton Creek, approximately 4.0 km southwest of the confluence of Barton Creek with the Colorado River (Figure 137). With 3.41 km of surveyed passages, it is presently the longest cave in the area and the sixth longest cave in Texas (Figure 138). The end-to-end extent of the cave is 980 m, one of the largest in this regard among the long caves of Texas (Appendix C).

The cave was discovered in 1971 through excavation of a low crawlway opening onto Barton Creek. Exploration and mapping began in that year and continued through 1974. The map of the cave (Figure 138) shows all presently mapped passages, but many branching passages remain unexplored or unmapped (Russell, 1975b).

Airman's Cave appears somewhat complex in plan-view (Figure 138); however, nearly the entire cave has formed within a 1 to 2 m-thick bed in the upper part of the Edwards Limestone. The most striking aspect of the cave map is extensive development of linear segments, suggesting strong structural control.

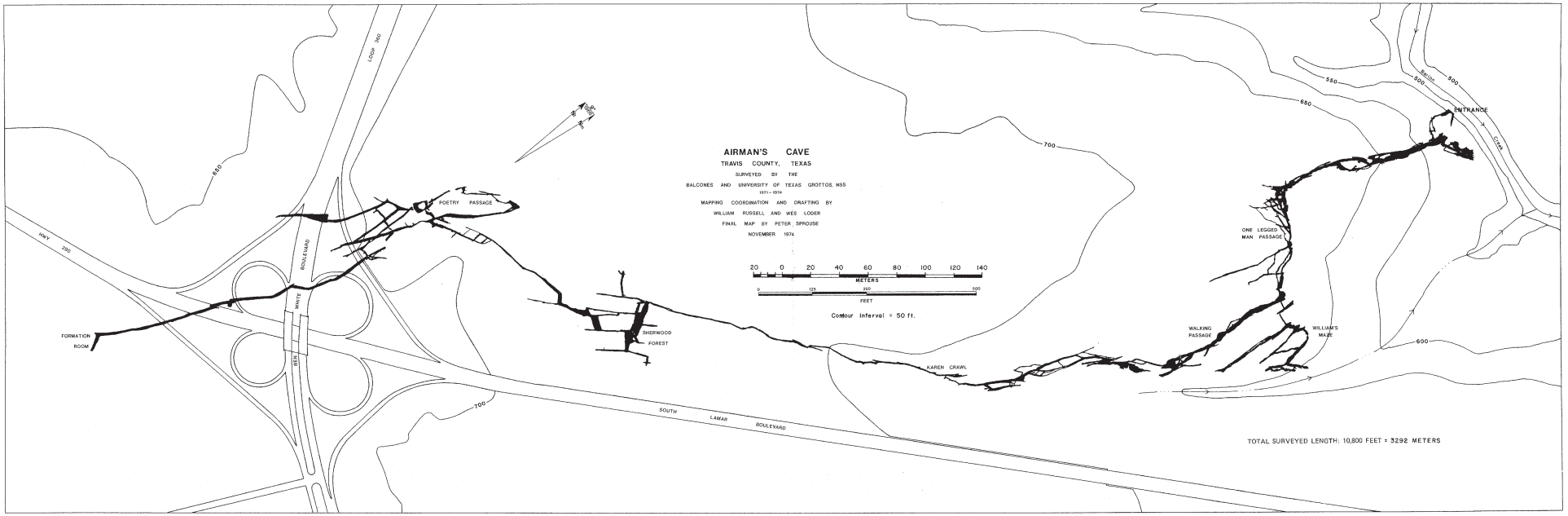


Figure 138: Map of Airman's Cave, Travis County. Modified from Russell (1975b, p. 176).

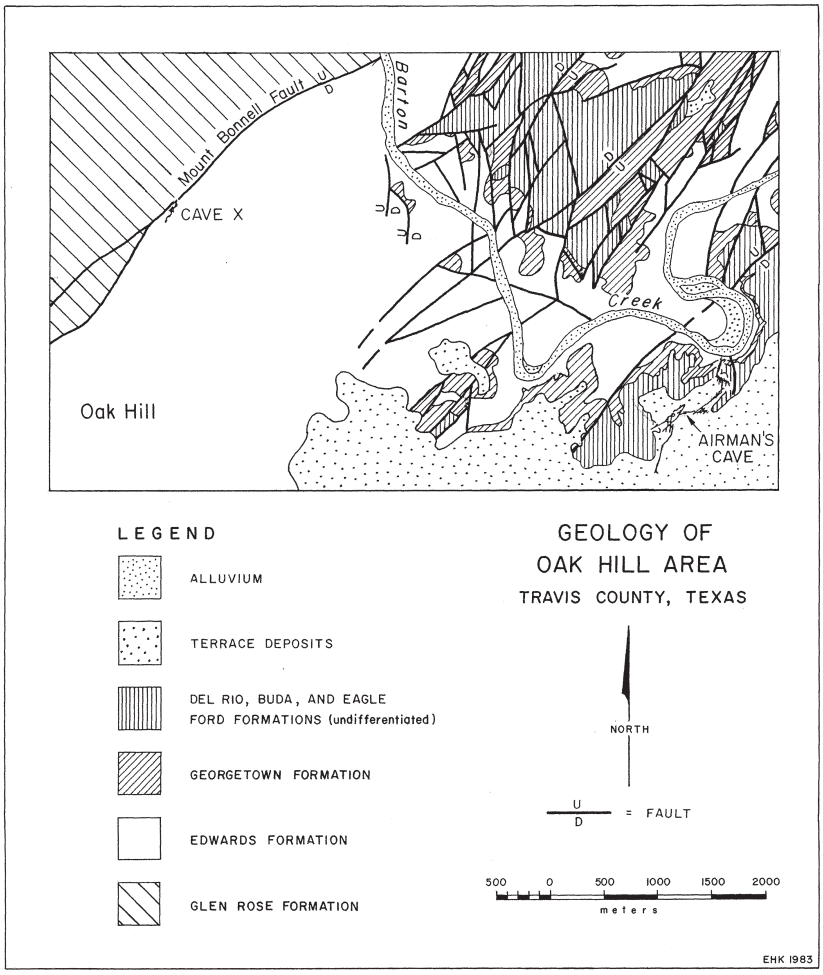


Figure 137: Geologic map of the vicinity of Airman's Cave and Cave X, Travis County. Based on maps by Rodda and others (1970) and Garner and Young (1976).

Figure 139: Typical passage in Airman's Cave. This view is looking south in a passage approximately 300 m from the entrance. Note joints in the ceiling.



The only entrance to the cave is at its northeastern end, where it intersects the valley of Barton Creek. Most passages of the cave are less than 1.5 m high, necessitating crawling and making exploration and mapping difficult. The Walking Passage, 100 m in length and beginning 230 m within the cave, is one of a few places high enough to stand erect. Much of the long passage leading to the Formation Room at the southwestern end of the cave is also this high.

The typical passage in Airman's Cave is 1 to 3 m in width and rectangular in cross-section (Figure 139). Joints are prominently visible in the ceiling and in places along the walls.

Airman's Cave contains no active streams, and most passages are very dry and free of mud. However, severe floods occasionally occur along streams in the Austin area (Breeding and Dalrymple, 1944) and Airman's Cave may receive floodwater from Barton Creek during such events. Speleothems are uncommon except in the southwestern end of the cave where flowstone, stalactites, and stalagmites are actively forming.

Baselevel in the vicinity of Airman's Cave is Barton Creek. The cave underlies a relatively nondissected surface that slopes gently at 5° toward Barton Creek

Cave X

Cave X is in the southwestern part of the Austin West U.S.G.S. 7.5-minute quadrangle, approximately 2.5 km northeast of Oak Hill (Figure 137). It lies about 300 m east of the Mount Bonnell Fault as mapped by Rodda and others (1970). The cave has been described by Reddell and Russell (1961a, p. 9-11), Russell (1974), and Fieseler and others (1978, p. 22-23).

Cave X has approximately 336 m of passages and is 193 m in extent (Figure 140). A single entrance at its southern end drops 4 m into a 2 m-high room. The cave extends to the northeast as a series of crawlways at progressively lower levels that are connected by pits varying from 3 to 6 m in depth (see profile, Figure 140). The lowest, explored part of the cave is the Los Pollitos Room at the northern end of the cave, 34 m below the entrance.

Passages leading to the First Big Room average 2 to 3 m in width and 1 m in height. The First Big Room has a similar ceiling height; however, it and passages leading from it are up to 6 m wide. Approximately 25 m beyond the First Big Room a 6 m drop leads to a linear passage 43 m long and 1 to 3 m high. This makes a right-angle bend to the northwest along a 1.5 m-wide, 0.6 m-high crawlway to a 6 m drop into the final room which is 17 m long, 9 m wide, and 10 m high.

The cave developed within porous beds of Member 2 of the Edwards Formation as mapped by Rodda and others (1970). This unit consists of hard, dense, argillaceous limestone that is flaggy at the top and nodular and cherty in the bottom third of the member. The lowest levels of the cave may lie at the top of Member 1 of the Edwards Formation at the former position of the Kirschberg Evaporite, which is locally present in the Austin area as a cavernous zone of solution-collapse (Rodda and others, 1970).

Runoff entering the cave during intense storms flows down the entrance passage and into Mystery Hole, a 6 m pit in the floor of this passage, 35 m south of the First Big Room (Reddell and Russell, 1961). Recharge from surficial runoff has been observed at several points in the area above the cave, and stagnant pools have been encountered in parts of the cave (Reddell and Russell, 1961a). Local baselevel is Barton Creek approximately 67 meters below and 2.4 km northeast of the entrance of Cave X.

Bee Creek Cave

Bee Creek Cave is 0.2 km northwest of the confluence of Bee Creek and the Colorado River in Austin, Travis County, in the east-central part of the Austin West U.S.G.S. 7.5-minute quadrangle. It opens to the surface at the base of a 3 m-high cliff near the top of the valley of Bee Creek. The cave is 38 m long and consists of a single 32 m-long passage with a 6-m segment on an upper level (Figure 141). The cave trends nearly straight at N 42° E. Height of the passage varies from 1.2 m at the entrance to 0.8 m midway

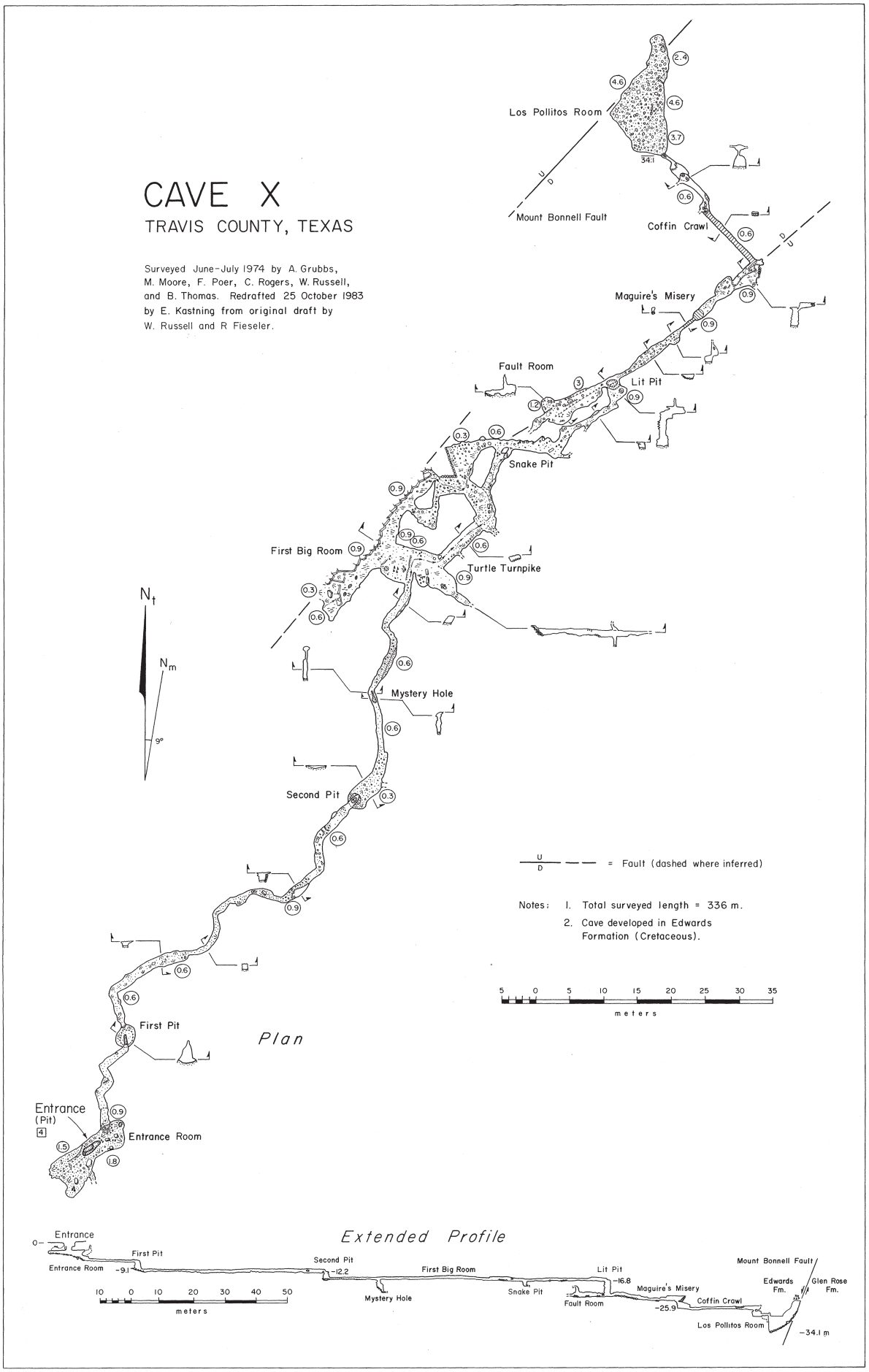


Figure 140: Map of Cave X, Travis County. Modified from Russell (1974, p. 152-153).

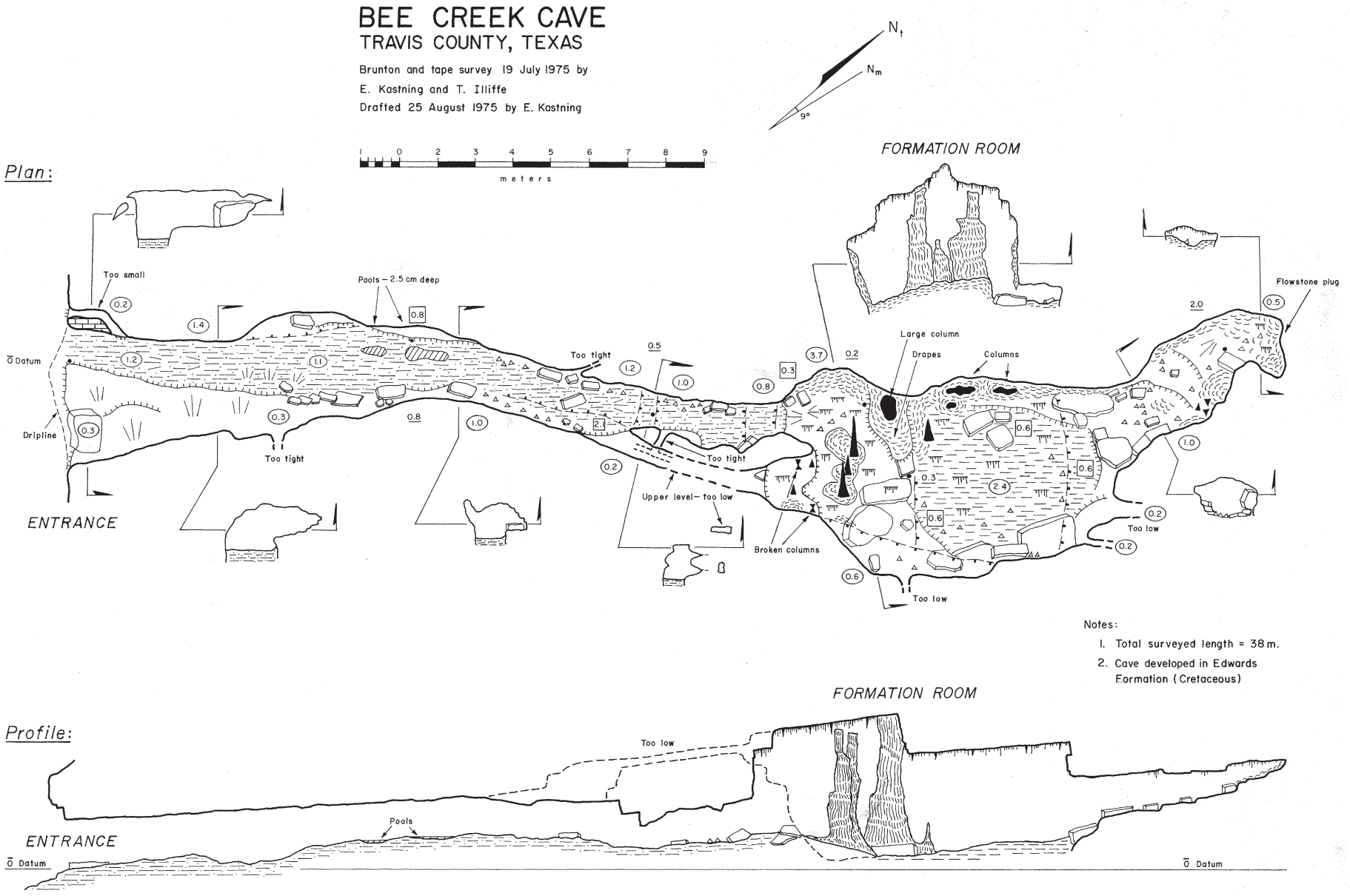


Figure 141: Map of Bee Creek Cave, Travis County.

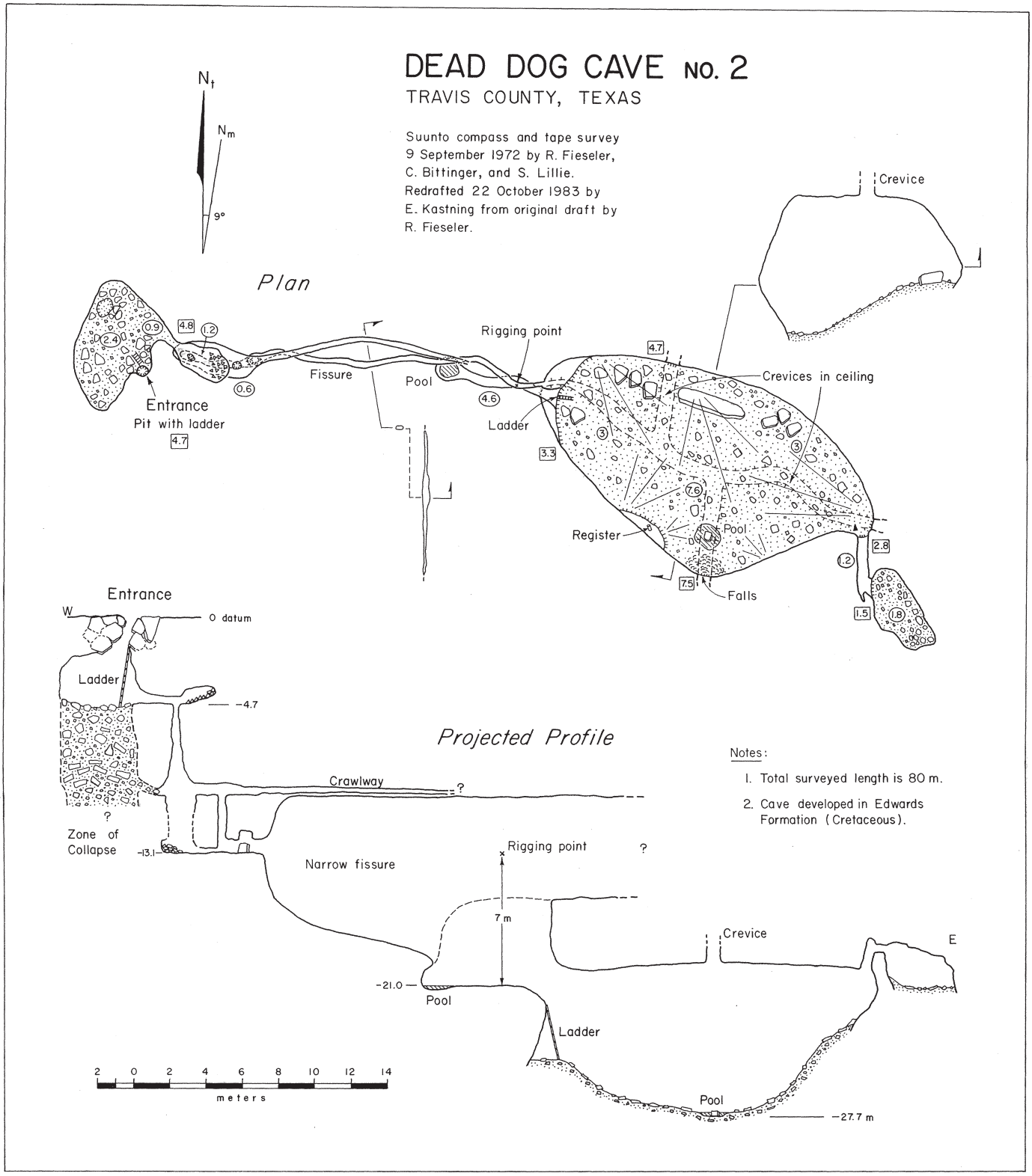


Figure 142: Map of Dead Dog Cave No. 2, Travis County. Modified from Fieseler (1976, p. 210).

along its length and 3.7 m in the Formation Room. A short crawlway leading from this room terminates in a flowstone-covered slope of debris.

Bee Creek Cave is entirely developed within a soft, nodular, burrowed member of the Edwards Formation (Member 3 of the Edwards as mapped by Rodda and others, 1970). Water enters the cave as seepage along joints; and, at times, small, shallow pools are found about 6 m inside the cave. The Formation Room contains massive columns and stalagmites up to 3.6 m high, flowstone, and small stalactites. They are actively forming because the cave receives seepage along joints from infiltration at the surface, only 5 m above the cave. Water entering the cave is lost through percolation into the floor of silt and clay and through evaporation.

The cave is about 250 m east of Mount Bonnell Fault, the major fault of the area, and within a few meters of two other northeasterly trending faults that merge with Mount Bonnell Fault (Rodda and others, 1970).

Dead Dog Cave No. 2

Dead Dog Cave No. 2 is 120 m west of the intersection of Balcones Drive (Mopac Highway) and Steck Avenue in the city of Austin (northwest corner of the Austin East U.S.G.S. 7.5-minute quadrangle). The cave has a mapped length of 80 m and depth of 28 m (Figure 142).

The entrance is a 5 m pit dropping into a 3 m-high room. To the east a pit drops 8 m further to a narrow, fissure-like passage that continues east for 21 m to where it intersects a large room below. This room is 20 m long, 11 m wide, and 8 m high. The fissure-like passage extends across the top of the room. A short passage, 4 m long, extends from the eastern end of the room and leads into a final, small chamber 6 m long, 3 m wide, and 2 m high. A 15 m-long crawlway running parallel to and above the fissure-like passage at the western end of the cave is the only other known passage.

The cave is only 120 m west of the Mount Bonnell Fault. The upper levels of the cave formed in Member 1 of the Edwards Formation as mapped to the west of the cave by Rodda and others (1970). This consists of burrowed, dolomitic limestone at the base, which is overlain by cherty, thin- to thick-bedded limestone, dolomite, and dolomitic limestone. The large lower room developed near the contact of the Cedar Park Limestone and Bee Cave Marl Members of the Walnut Formation, where the beds consist of extensively burrowed, nodular, fine-grained limestone. The contact of the Bee Cave Member with the underlying Bull Creek Member should be no more than a few meters below the floor of the lower room because locally the Bee Cave Member is only 9 m thick (Moore, 1964; Rodda and others, 1970).

Dead Dog Cave No. 2 contains a few pools of water, and seepage enters along the southern wall of the lower room as a small waterfall dripping from a prominent joint in the ceiling. There are some speleothems in the cave. Shoal Creek, approximately 570 m to the east and 17 m below the entrance of the cave, is the local baselevel for surficial runoff.

Swimmin' Hole

Swimmin' Hole is a small cave 2.3 km north of Cat Mountain and 0.5 km southeast of Bull Creek in northwest Austin (northeast Austin West U.S.G.S. 7.5-minute quadrangle). It is 50 m southeast of the intersection of Valburn Drive and Valburn Circle.

This cave is 13 m long and averages 3 m in width and 1.5 m in height (Figure 143). It was discovered on 9 January 1979 during excavation for a swimming pool in a residential neighborhood and explored and mapped during the course of the present research. Most of the cave was subsequently destroyed by further excavation and installation of the pool. The cave did not have a natural entrance.

Swimmin' Hole developed within a highly burrowed bed of dolomitic limestone in the lowermost part of Member 1 of the Edwards Formation (Rodda and others, 1970). A hard, crystalline, dolomitic bed lies between the ceiling of the cave and the surface, a vertical distance of 1.5 m. This gives the cave a stable roof and has prevented collapse. During excavation for the pool another cave of unknown dimensions was intersected along the same strata southeast of Swimmin' Hole but was sealed before it could be explored. Two core-holes drilled by a local,

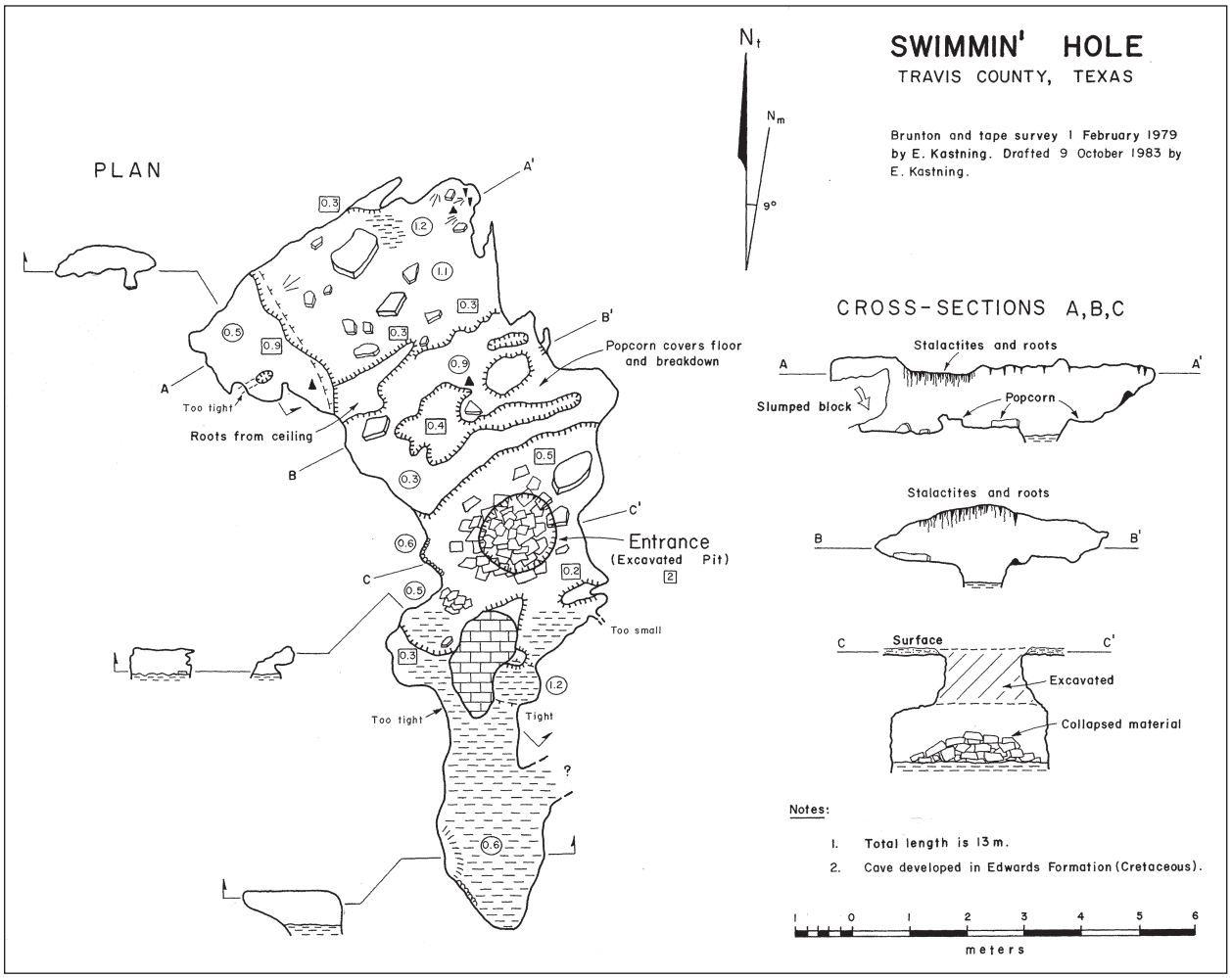


Figure 143: Map of Swimmin' Hole, Travis County

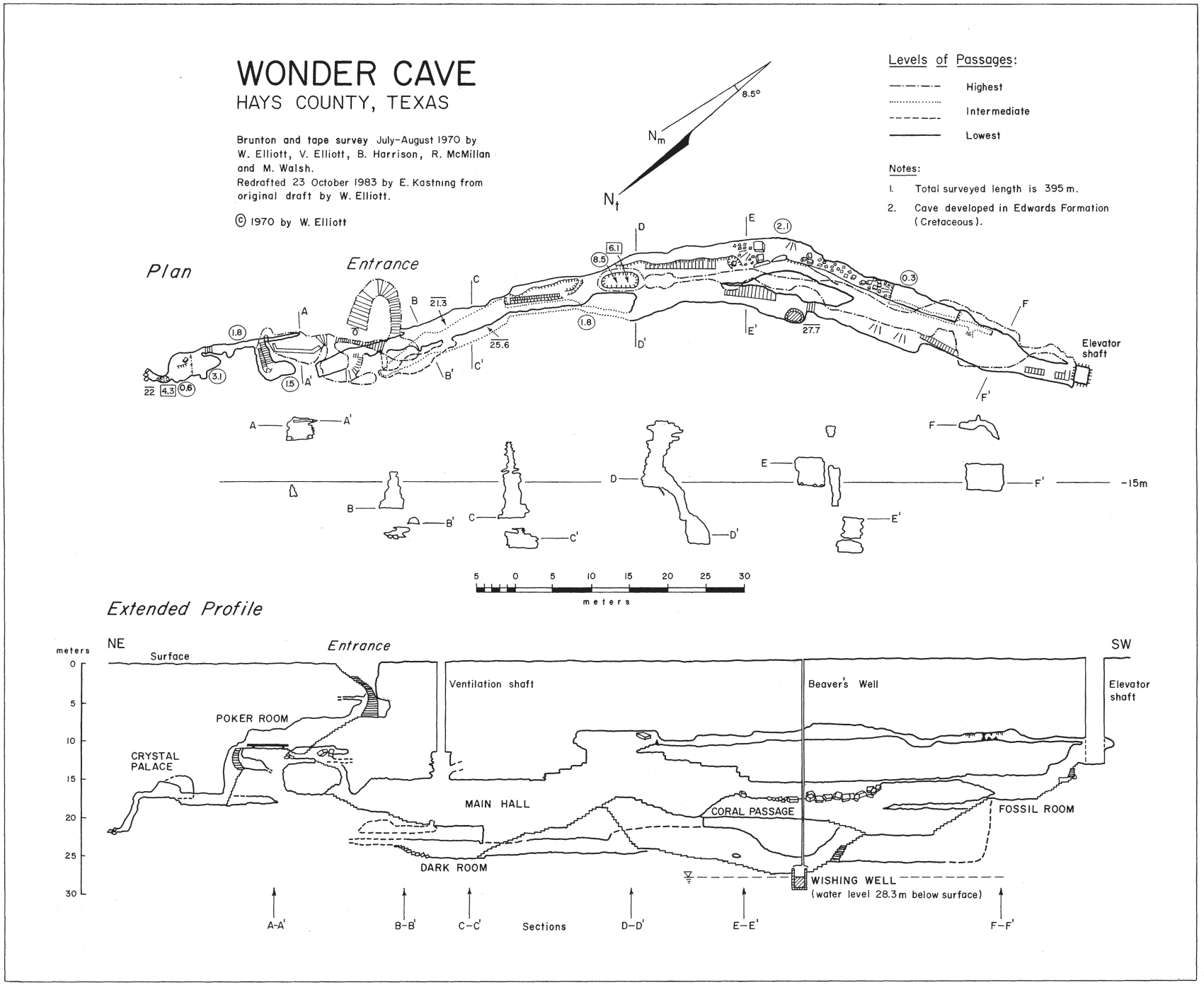


Figure 144: Map of Wonder Cave, Hays County. Modified from Elliott (1970a, p. 193).

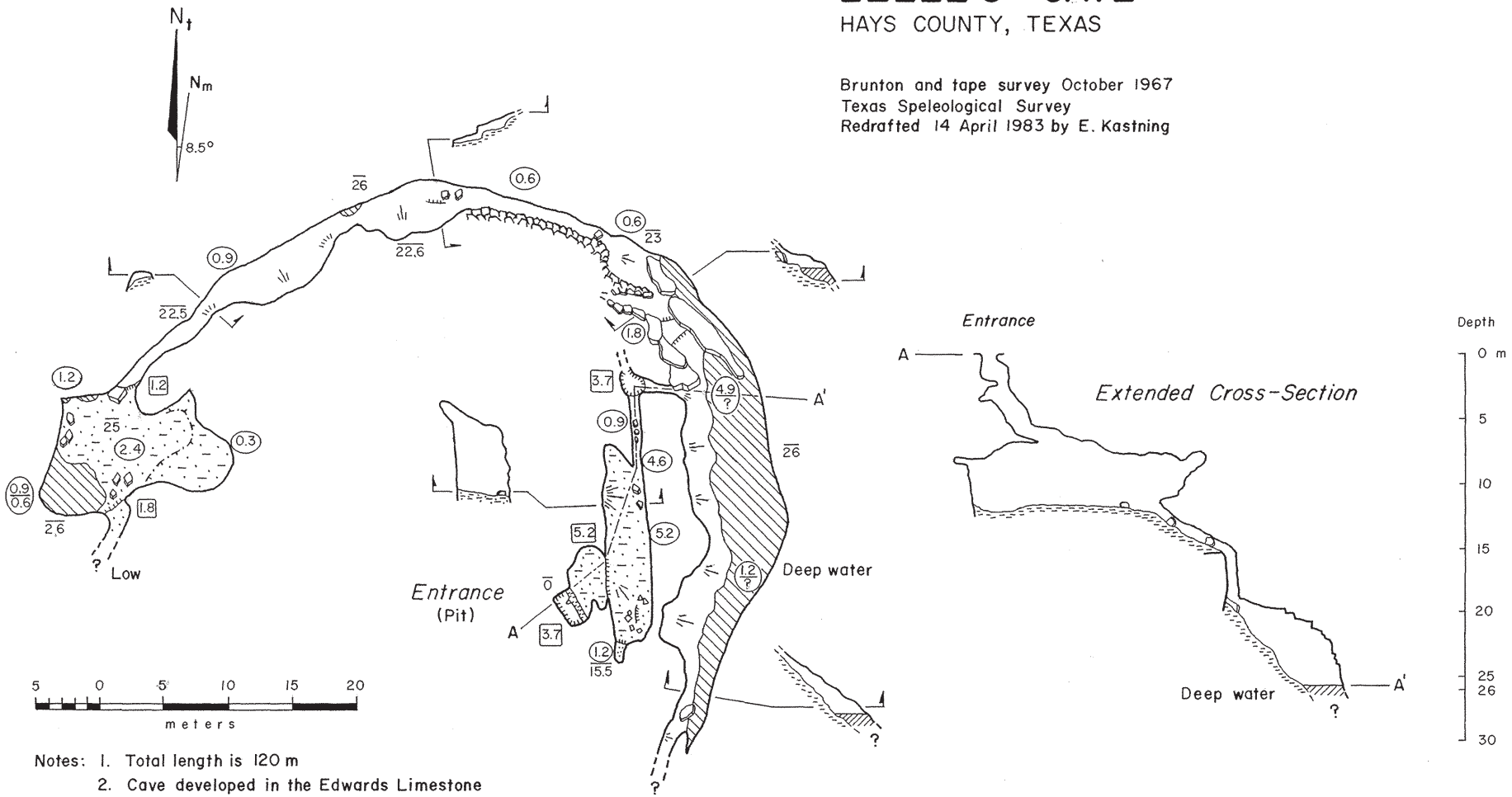


Figure 145: Map of Ezell's Cave, Hays County. Modified from Russell (1976, p. 21).

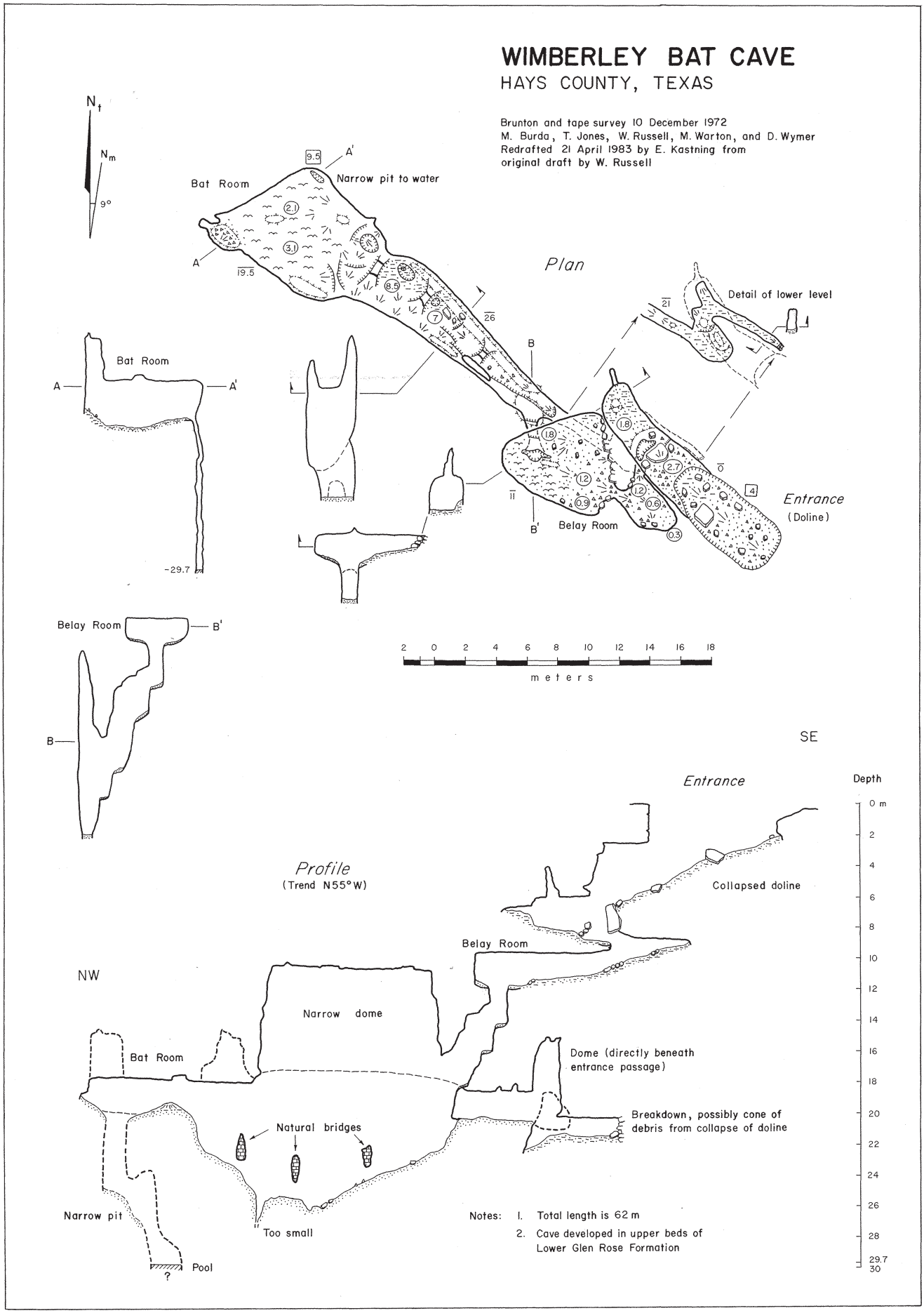


Figure 146: Map of Wimberley Bat Cave, Hays County. Modified from Russell (1973, p. 142-143).

engineering-consulting firm northeast and south of the cave intersected a zone of high permeability at the same horizon as the cave.

The cave was dry and devoid of standing pools. However, the ceiling, floor, and walls of the entire cave were covered with coralline "popcorn" indicative of wetter conditions. Stalactites up to 15 cm in length, but no longer active, occurred throughout the cave, and many had roots emerging through their hollow centers. The cave lies on top of a narrow ridge formed by Bull Creek, 102 m below on the north and west, and a small unnamed tributary to it 47 m below on the east.

Wonder Cave

Wonder Cave is southeast of Prospect Street within the city limits of San Marcos, 1.3 km southeast of the intersection of Routes 12 and 2439 (San Marcos North U.S.G.S. 7.5-minute quadrangle). The cave has developed along several levels superimposed on one another (Figure 144). Although it is confined to a lateral extent of 120 m, approximately 395 m of survey lines were used to accurately map it (Elliott, 1970a). The lowest point in the cave, at the Wishing Well, is 28 m below the surface.

There are two entrances to the cave. One is a collapsed doline in the northeastern part of the cave, and the other is an artificial elevator shaft at the southeastern end of the cave. Originally known as Beaver Cave, Wonder Cave was opened to the public around 1908 as Texas' first commercial cave (Craun, 1948; Russell, 1976).

The cave follows three main trends: N 61.5° E in the southern end, N 42° E in the northern end, and N 27° E in the central part (Figure 144). The general trend of the cave is roughly parallel to the San Marcos Springs Fault. The cave developed in the Edwards Limestone, and various levels occupy highly burrowed strata of enhanced porosity.

Wonder Cave does not contain any active streams. However, the local piezometric surface is intersected by the Wishing Well at 28.4 m below the surface. Speleothems are sparse except for flowstone and stalactites adorning the Crystal Palace at the northeastern end of the cave.

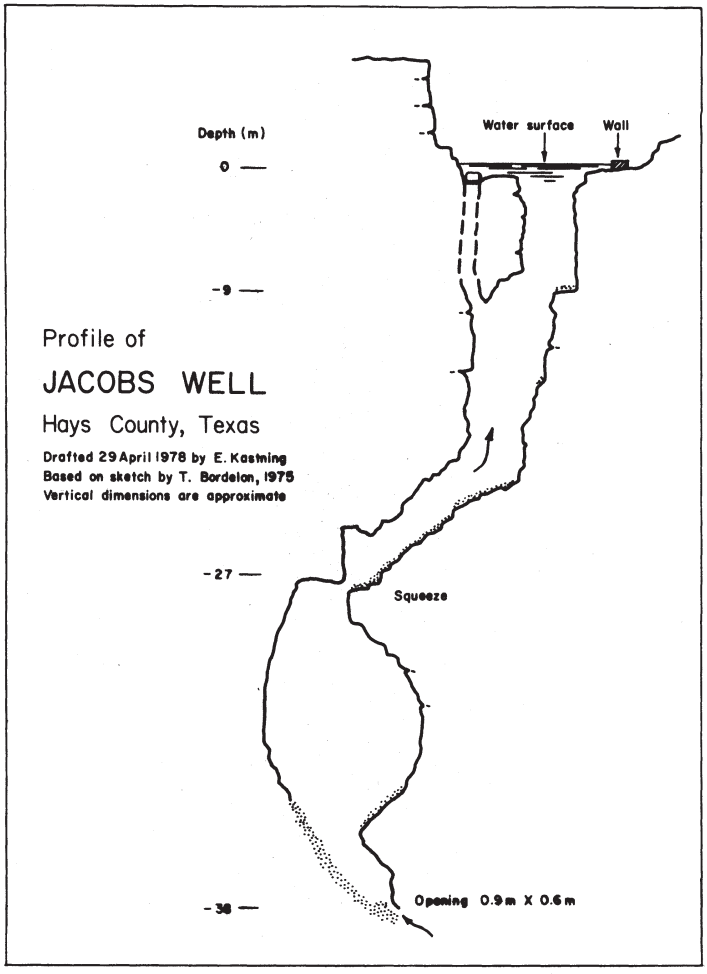


Figure 147: Profile of Jacobs Well, Hays County. Modified from Bordelon (1975, p. 39) and Kastning (1978, p. 15)

Wonder Cave lies on the slope of the Balcones Escarpment, and local surficial drainage is to the southeast to Purgatory Creek, 0.8 km away and 24 m lower. The cave lies within a faulted block bounded on the southeast by the San Marcos Springs Fault, 0.2 km away, and on the north by another fault, 0.3 km away (Grimshaw, 1976; Russell, 1976).

Ezell's Cave

Ezell's Cave is approximately 0.5 km southwest of Wonder Cave in the north-central part of the San Marcos South U.S.G.S. 7.5-minute quadrangle. The cave has 120 m of surveyed passage and reaches a depth of greater than 26 m below the entrance (Figure 145).

The entrance consists of a 12 m-deep pit leading to a room 15 m long, 4 m wide, and 5 m high from which a 1 m-high crawl leads 6 m to a 4 m drop to the main passage. This passage is arcuate in plan-view and consists of two rooms connected by a 1 m-high crawlway. The northwestern room is 2.5 m high and contains a small pool 0.6 m deep. The room along the eastern side of the cave is 1.5 to 5 m in height and contains a long pool of undetermined depth. The cave developed within the Edwards Limestone in strata similar to that containing Wonder Cave.

Breakdown forms the inner wall of the arcuate passage and bedrock forms the outer wall. Ceilings and floors of passages slope away from the center of the arc.

The level of water in Ezell's Cave is approximately at the same elevation as that in the Wishing Well of Wonder Cave (Craun, 1948; Russell, 1976), suggesting a connection in groundwater between them. Levels of water typically fluctuate one to two meters over a year, and flakes of calcite are commonly found floating on the surface of the water, suggesting that the water is supersaturated with respect to calcite (Russell, 1976).

The cave has received widespread publicity since its discovery around 1870, because it has harbored a large population of the Texas blind salamander, *Eurycea (Typhlomolge) rathbuni*. Owing to vandalism, pollution, and indiscriminant collection, specimens are rarely seen in the cave, despite recent measures of conservation (Davis, 1971).

Like Wonder Cave Ezell's Cave lies on the flank of the Balcones Escarpment. Local drainage is south to Purgatory Creek, 300 m away and 21 m lower.

Wimberley Bat Cave

Wimberley Bat Cave is 6.4 km northwest of Wimberley in Hays County (Rough Hollow U.S.G.S. 7.5-minute quadrangle). It lies east of Cypress Creek and about 1.6 km upstream from Jacobs Well. Total surveyed length of the cave is 62 m, and it reaches a depth of 29.7 m below the entrance (Figure 146). Horizontal extent of the cave is 37 m.

The entrance to Wimberley Bat Cave is in a collapsed doline with a floor sloping into the first of a series of three rooms. These are connected by pits to the Bat Room (21 m long, 3 to 6 m wide, and 2 to 6 m high) forming the lower level of the cave. The cave developed on four distinct levels at 8, 11, 20, and 25 m below the surface. At the northwestern end of the Bat Room a narrow pit drops 9.5 m to the level of water at the deepest point in the cave. Passages developed along favorable strata of the lower member of the Glen Rose Formation. In the vicinity of Wimberley Bat Cave the upper beds of the lower Glen Rose are within a reefal facies (Grimshaw, 1970; Russell, 1973).

Wimberley Bat Cave does not contain any streams, and the only water in the cave is at the lowest point. Local, surficial drainage is toward Cypress Creek to the west.

Jacobs Well

Jacobs Well is in the bed of Cypress Creek, 2.3 km northwest of Wimberley. It is an artesian spring issuing from a nearly vertical conduit at least 38 m deep (Figure 147). The width of the opening at the surface is about 3 m (Figure 148), and the passage widens downward into chambers connected by constrictions. One narrow point 28 m down is particularly hazardous to divers as this occasionally becomes choked with sediment. It has recently been sealed, following the death of a diver.

The cave developed in the cavernous lower member of the Glen Rose Formation. Jacobs Well Fault, approximately 280 m downstream, has placed marls and thin limestones in contact with the cavernous zone. Groundwater flowing in the subsurface is forced upward to the surface by this impermeable barrier (Kastning, 1978, p. 14-15). The water does not flow to the surface along the plane of the fault, but rises along fractures associated with the fault that provided a channel for flow early in the history of the spring (Grimshaw, 1970). Discharge from Jacobs Well varies from 68 to 170 liters per second (DeCook, 1963; Brune, 1975).

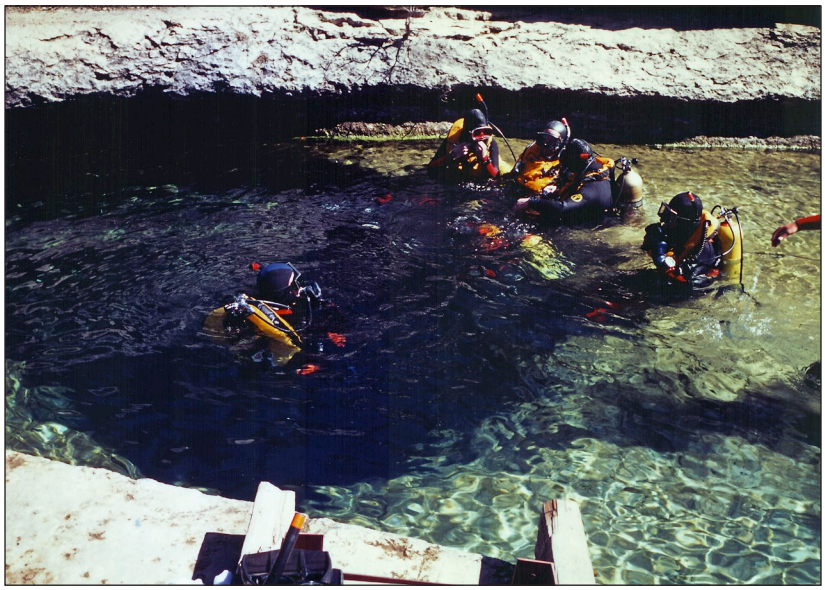


Figure 148: Divers at the entrance to Jacobs Well.

Speleogenesis

Comparisons among the mine caves described above show various similarities in the morphology of passages and suggest that some controls on development of the caves were similar. However, the caves differ markedly in some aspects.

Lithologic Control

Each cave, with the exception of Jacobs Well, exhibits strong, lithologic control as suggested by passages conformal to stratigraphic horizons. Porous, burrowed zones in the Edwards Limestone have developed into the most cavernous horizons. In some examples, such as in Airman's Cave, Cave X, and Dead Dog Cave No. 2, primary porosity derived from burrows may have been only slight; but the contrast with denser, overlying and underlying rocks appears to have been sufficient to promote the flow of groundwater within specific beds. In other caves, including Swimmin' Hole and Wonder Cave, the initial porosity of the bedrock was much greater. In Wonder Cave vertical changes in lithology are apparent from cross-sections of the cave (Figure 144). Beds of higher porosity or greater content of calcite are recessive in contrast to those that are denser or more dolomitic. Differential solution among beds of varying lithology has resulted in undercutting of less soluble rocks and collapse of large blocks of breakdown within the cave.

Structural Control

Each cave consists of well-defined, linear segments of passages. Observations of fractures in the ceilings and walls indicate that joints have strongly guided the orientations of developing passages. This is well demonstrated by rose-diagrams in the case of Airman's Cave and Cave X (Figure 149). Many fractures in Airman's Cave, Cave X, Bee Creek Cave, Dead Dog Cave No. 2, and Wonder Cave are parallel to faults of the Balcones Fault Zone. Trends of fractures in Airman's Cave and Cave X agree well with those mapped on the surface by Muehlberger and Kurie (1956, p. 47). Minor faults were observed at the entrance to Airman's Cave and in the Poetry Passage (Russell, 1975b). Others occur in the Fault and Los Pollitos Rooms of Cave X, and in Wonder Cave. Wonder Cave has formed along a zone of fractures associated with a minor fault of the Balcones Fault Zone, where displacement has been about 1 m (Russell, 1976).

Collapse has modified Wonder, Wimberley Bat, and Ezell's Caves and perhaps even Cave X and Dead Dog Cave No. 2. In Wonder Cave collapse was induced by differential solution as described above, but in Ezell's Cave the process was similar to that described for Natural Bridge Caverns (Chapter 10). Ezell's Cave follows the flank of a cone of debris as indicated by cross-sections of sloping passages, arcuate shape of the cave in plan-view, and abundant breakdown along the inside wall of the arc (Figure 145). Apparently the cave was once a larger chamber at lower depth than that of the present cave, and, as solution progressed, the void increased to a critical size where beds of the ceiling became unstable and lead to collapse and upward stoping of passages. The cave stabilized in its present configuration in a manner similar to that forming the Hall of the Mountain Kings and the doline at the entrance of Natural Bridge Caverns (Figure 135). Locally, the piezometric surface in Ezell's Cave is now 26 m below the surface, intersecting the cone of debris part way up, and may reflect ponding caused by collapse.

Hydrogeologic Control

All caves described in this chapter are associated with the Balcones Fault Zone. The hydrogeology of the Austin-San Marcos area was complicated by numerous faults that have locally compartmentalized groundwater. The distribution of the troglobitic faunas suggests that the flowpaths of groundwater have been locally confined within faulted blocks. Uhlenhuth (1921, 1948) and Russell (1976) have suggested a hydrologic communication among caves and wells near Purgatory Creek, Ezell's Cave, Wonder Cave, the artesian well of the U.S. Fish Hatchery at San Marcos, and San Marcos Springs. Groundwater apparently moves northeastwardly within blocks defined by the extensive San Marcos Springs and Comal Springs Faults (Guyton and Associates, 1979).

Bee Creek Cave, Swimmin' Hole, and Dead Dog Cave No. 2 are relatively short caves consisting of poorly integrated solutional cavities. Bee Creek Cave developed in the phreatic zone along a prominent fracture associated with the Mount Bonnell Fault. It is presently well above local baselevel and is a relict cave associated with incision of Bee Creek and the Colorado River shortly after Balcones faulting. Swimmin' Hole, a relict cave located well above Bull Creek, developed within a porous bed lying between dense resistant units. It exhibits minor control by fractures. Dead Dog Cave No. 2 consists of two solutional cavities connected along an east-west fracture. One is developed in the lower most Edwards Limestone and the other in underlying beds of the Walnut Formation. The development of this cave is probably more recent than that of Bee Creek Cave and Swimmin' Hole as a result of its lower topographic position. It is also somewhat better integrated than the other two.

Airman's Cave shows evidence of extensive development and modification by floodwater. There are three maze-like sections in the cave (Figure 138): one extending from the entrance to Karen Crawl, a second at Sherwood Forest, and a third in the vicinity of the Poetry Passage. The greater abundance of maze-like passages near Barton Creek and the abruptly terminating, joint-controlled passages indicate that these mazes were primarily excavated by floodwater (Palmer, 1975).

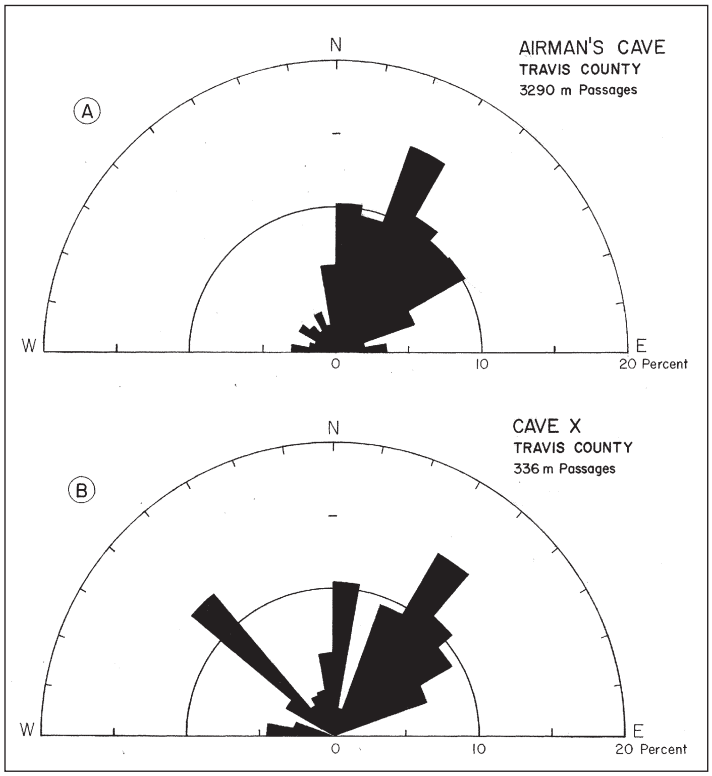


Figure 149: Rose-diagrams of linear segments of passages in (A) Airman's Cave and (B) Cave X as measured from maps of Figures 138 and 140 respectively.

Maze-like passages near the entrance to Airman's Cave were probably excavated by backflooding from Barton Creek at a time when the creekbed was at a higher elevation. Mazes in Sherwood Forest and in the southern part of the cave may have originated in response to periodic flooding as northwardly-flowing, vadose water ponded behind constrictions during high runoff. Passages in Sherwood Forest have enlarged along bedding-plane partings and joints. Maze-like passages at the southern end of the cave may have developed as diversionary routes for floodwater around constrictions. Although not documented in this study, it is likely that the cave occasionally floods at the present time during high discharge along Barton Creek.

Chronology of Speleogenesis

Airmans Cave is the best integrated cave of those studied in the Austin-San Marcos area. The chronology of its development may be summarized as follows:

1. Incision of the Colorado River and Barton Creek may have begun in response to uplift of the Edwards Plateau in the Miocene Period. However, incision was accelerated upon rejuvenation of the planed surface during the middle to late Pleistocene. Flow of groundwater increasedalong fractures and bedding-plane partings as baselevels were progressively lowered.

2. Most fracture in the area originated from tensional forces during Balcones faulting and were thereby sufficiently open to easily conduct groundwater from points of recharge north to Barton Creek. A single, major conduit developed within the shallow-phreatic zone along the most open fractures.

3. Velocities of flow through the conduit increased as incision of Barton Creek and excavation of the cave continued. This in turn accelerated enlargement of the conduit. Some of the tributary-passages in the southern end of the cave may have developed at this time as adjacent flowpaths became integrated with the master conduit.

4. Further incision of Barton Creek resulted in a drop in the piezometric surface below the level of Airman's Cave. The cave was then within the zone of flooding and subjected to considerable backflooding during periods of high discharge on Barton Creek. Chemically and hydrodynamically aggressive floodwater enlarged most of the passages between the entrance and Karen Crawl. Many fractures intersecting the master conduit were enlarged at this time, forming numerous, short, side passages and giving sections of the cave a maze-like character

5. Increased recharge in the area of catchment draining through Airman's Cave caused the rate of discharge through the cave to exceed its capacity to transmit the flow. Mazes developed in the southern parts of the cave where water ponded behind constrictions. Increased hydraulic gradients diverted floodwater around blockages in the Poetry Passage, resulting in a network-type of maze.

6. There is some evidence suggesting that channels of diversion developed in the vicinity of the Walking Passage and William's Maze. The initial conduit may have extended from Karen Crawl to the area that is now the southern end of the Walking Passage, and then toward the northeasternmost passage of William's Maze and northwest toward the One Legged Man Passage. Progressive routes of diversion, each further toward the west and larger in cross-section, may have formed in response to flooding. Ultimately the Walking Passage became the dominant route and enlarged to its present dimensions. In effect, therefore, the parallel passages of William's Maze may have been successive meander-cutoffs. This mechanism was originally suggested by Russell (1975b). The validity of such an interpretation may be tested once the cave is surveyed with precise vertical control; however, this was not attempted during the present study.

7. Further incision of Barton Creek has left Airman's Cave above the zone of frequent flooding. Although no floods along Barton Creek have been observed to rise to Airman's Cave, located 6 to 7 m above the bed of the creek, the cave may possibly be within the zone of flooding during discharges of high magnitude.

Several caves in the Austin-San Marcos area consist of distinct levels. Even though excavated passages are strongly positioned along favorable lithostratigraphic horizons, some levels may reflect periods of enlargement wherein passages evolved in a descending sequence in response to changes in local baselevel. Examination of the maps of Cave X, Dead Dog Cave No. 2, Wonder Cave, and Wimberley Bat Cave (Figures 140, 142, 144, and 146) suggests that levels may have been thus guided by topographic changes. However, this mechanism, although as plausible in this area as elsewhere on the Edwards Plateau, is difficult to substantiate. In any case, most, if not all, caves discussed in this chapter developed after the middle of the Pleistocene in response to accelerated dissection of the uplands by surficial streams. Most caves are positioned along fractures associated with uplift and faulting that began in the Miocene.

SUMMARY

Landforms in the Austin-San Marcos area, like those in the New Braunfels area (Chapter 10) and Georgetown area (Chapter 12), were strongly influenced by the Balcones Fault Zone. Many passages of caves are oriented along fractures produced by faulting, and collapsed chambers are positioned where the density of fractures is particularly high. The complexity and intensity of faulting has partitioned the flow of groundwater within northeast-trending fault-blocks. During speleogenesis, as well as before and since, groundwater moved parallel to the Balcones trend to outlets in valleys or upward along fractures under artesian conditions to flowing springs. Some caves remaining close to baselevel were later modified by periodic flooding. Others, where surficial incision has been deep, have long since been drained and remain as relict caves in the uplands, considerably above present-day streams.

12
GEORGETOWN AREA

DESCRIPTION OF AREA

The Georgetown area lies within the Balcones Fault Zone in west-central Williamson County in the northeastern part of the Edwards Plateau (Plate 1). It is bounded on the south and north along lines of 30°35'00" and 30°40'00"north latitude respectively and on the east and west along lines of 97°38'30" and 97°43'30" west latitude respectively. The area lies within the northwestern and southwestern 7.5-minute quadrants of the Round Rock 15-minute quadrangle (1:24,000-scale topographic maps have not been published at this writing), and is drained by the San Gabriel River, a major tributary of the Brazos River (Figure 150). The southeastern half of the area is principally drained by Smith Branch, a tributary of the San Gabriel River, whereas the northeastern half is drained by short creeks flowing directly into the North, Middle, and South Forks of the San Gabriel River. The city of Georgetown is located in the north-central part of the area, in the vicinity of the confluences of the three forks of the San Gabriel River.

The geology of the Georgetown area is shown on maps by Taff (1892, plate 10, scale 1:262,000), Wilson and Philpott (1949, figure 1, scale 1:195,000), Holcomb and Minton (1955, figure 1, scale 1:195,000), Ward (1950, plate 1, scale 1:21,120), Walls (1950, plate 1, scale 1:20,000), and Barnes (1974a, scale 1:250,000). Exposed rocks in the Georgetown area consist of the Edwards, Comanche Peak, and Walnut Formations (Lower Cretaceous) in the western half of the area and of the Georgetown Formation (Lower Cretaceous), Del Rio Clay, Buda Limestone, Eagle Ford Group, and Austin Chalk (Upper Cretaceous) in the eastern half (Figure 150). Geologic sections in the vicinity of the Georgetown area are described by Ward (1950), Walls (1950), Matthews (1951), Lozo (1959), Moore (1964), Rodda and others (1966), and Wilbert (1963, 1967).

The Edwards Formation consists of a sequence of beds of limestone, dolostone, and chert. Limestone is generally aphanitic to fine-grained, but some strata are medium- to coarse-grained. The limestone is massively to thinly bedded, hard, and brittle; in places it is nodular, cherty, or crystalline (Rodda and others, 1966; Ward, 1950; Barnes, 1974a). Fossiliferous zones contain abundant molluscs, including miliolids and rudists, generally within a biomicritic or biosparitic matrix. The abundance of rudistid corals and other fossils, composition, texture, and depositional and stratigraphic relations indicate that in the Georgetown vicinity and to the north the Edwards Formation may be largely reefal in origin (Matthews, 1951, 1956; Nelson, 1959, 1973; Rodda and others, 1966; Fisher and Rodda, 1967b, 1969; Rodda and Fisher, 1969). Dolostone of the Edwards Formation tends to be fine-grained, soft, massively to thinly bedded, moldic, and porous. Nodular chert and lenticular bedding is common. Chert as nodules and plates varies in amount among beds of limestone and dolostone. Weathered zones of the Edwards are considerably recrystallized and honeycombed, giving the rock a cavernous texture and locally high permeability as an aquifer. The thickness of the Edwards Formation in the vicinity of Georgetown is 18 to 50 m, thinning northwardly. The Walnut and Comanche Peak Formations crop out in the west-central part of the area. Here incision of the South Fork of the San Gabriel River has exposed these units which otherwise remain buried in the remainder of the area.

The Georgetown Formation, which covers most of the eastern half of the area (Figure 150), consists mostly of limestone, but it contains beds of marl and some shale. The limestone is generally soft, fine-grained, argillaceous, nodular, and moderately indurated. Some strata are relatively hard, brittle, and thinly bedded (Barnes, 1974a). Texturally, the beds of limestone are biomicritic. Because they are argillaceous, marly beds weather rapidly, causing slopes to recede quickly (Barnes, 1974a). Beds of limestone and some beds of marl of the Georgetown Formation contain burrows and casts of burrows. Irregular bedding in the Georgetown is in part the result of burrowing organisms. Abundant nodules of micrite occur in the burrowed and marly beds. Disseminated and nodular pyrite is common in the Georgetown, and most of the formation is also glauconitic. The Georgetown is approximately 32 m thick in the area, but it thins southwardly onto the San Marcos Arch in Travis County, generally in response to thinning of individual beds (Ward, 1950; Lozo, 1959; Wilbert, 1963, 1967).

The Del Rio Clay, Buda Limestone, Eagle Ford Group, and Austin Chalk are exposed in the southeastern part of the Georgetown area (Figure 150) and form an upland 15 to 45 m higher in elevation than the surface developed on the Edwards and Georgetown Formations to the north and west (Figure 151). These younger beds, which are described in Ward (1950), Walls (1950), Tydlaska (1951), Atchison (1954), Barnes (1974a), and Appendix B, have little speleogenetic significance in the Georgetown area. However, the Buda and Austin are relatively resistant to erosion in comparison to the Del Rio Clay and Georgetown Formation. This is particularly true of the hard, massive, bioclastic Buda Limestone which upholds the escarpment along the southeastern margin of the drainage basin of the West Fork of Smith Branch (Figure 150).

Regional dip in the Georgetown area is approximately 0.1° to 1.0° to the southeast (Ward, 1950, p. 40 and plate 1; Walls, 1950, p. 42 and plate 1; see also Figure 151). Locally, especially in proximity to major faults, dips may be as great as 2° to 5° (Ward, 1950, p. 41).

Several major, *en echelon* faults of the Balcones Fault Zone pass through the area (Link, 1929; Barnes, 1974a; see also Figure 150). The largest of these, trending N 10°-20° E, forms the boundary between the Edwards Formation in the west and younger rocks to the east. It passes through Georgetown and 370 m east of the entrance to Inner Space Cavern (Figure 150). Displacement along the fault varies from 38 m where it crosses the South Fork of the San Gabriel River to possibly as much as 65 m near the southern boundary of the area (Ward, 1950, p. 40-41). In the southern part of the area the fault bifurcates into two significant and parallel branches that can be followed southwardly for at least 8 km (Barnes, 1974a). The western fault passes 210 m west of the entrance to Inner Space Cavern. Both branches are downthrown to the southeast and form a sizeable faulted block which encloses the cave (Figure 150).

Lesser faults are common throughout the area and many branch from the larger faults indicated on Figure 150. Minor, normal faults of small displacement, that are exposed along the banks of the San Gabriel River, are typically parallel to the Balcones trend of faults and in some places form grabens and horsts (Ward, 1950, p. 40; Walls, 1950, p. 42-43).

Rocks of the Georgetown area are well-jointed with most joints oriented vertically and spaced a few meters apart. In general, fractures are either parallel or orthogonal to major faults, especially where they lie close to them (Evans, 1965; Rogers, 1963). Away from large faults, the frequency of fractures decreases and orientations of fractures become less systematic.

Both lithology and geologic structure have controlled erosional development in the Georgetown area. Soft, marly units, such as the Georgetown Formation have eroded to gentle (0.1° to 0.5°) slopes east of the major fault; whereas hard, dense carbonate rocks, such as the Edwards Formation and Buda Limestone form steeper (1.0° to 6.0°) slopes west of the fault. This is particularly evident from topographic maps, where walls of valleys of streams are relatively steep in the Edwards Formation and Buda Limestone, but comparatively gentle in the outcrop of the Georgetown.

Faulting has modified the configuration of surficial drainage in the area. Streams have been deflected at or near faults and commonly follow traces of faults and associated fractures. This is apparent in segments of streams of the South Fork of the San Gabriel River and Smith Branch (Figure 150). Upstream tributaries of the West Fork of Smith Branch are either nearly coincident with major faults or are oriented parallel or perpendicular to them, suggesting structural control.

DISTRIBUTION OF CAVES AND PREVIOUS INVESTIGATIONS

Caves in Williamson County are principally distributed along the outcrop of the Edwards Formation, a 6 to 17 km-wide band extending north-south through the county just west of Georgetown. Most caves are within 17 km of Georgetown,

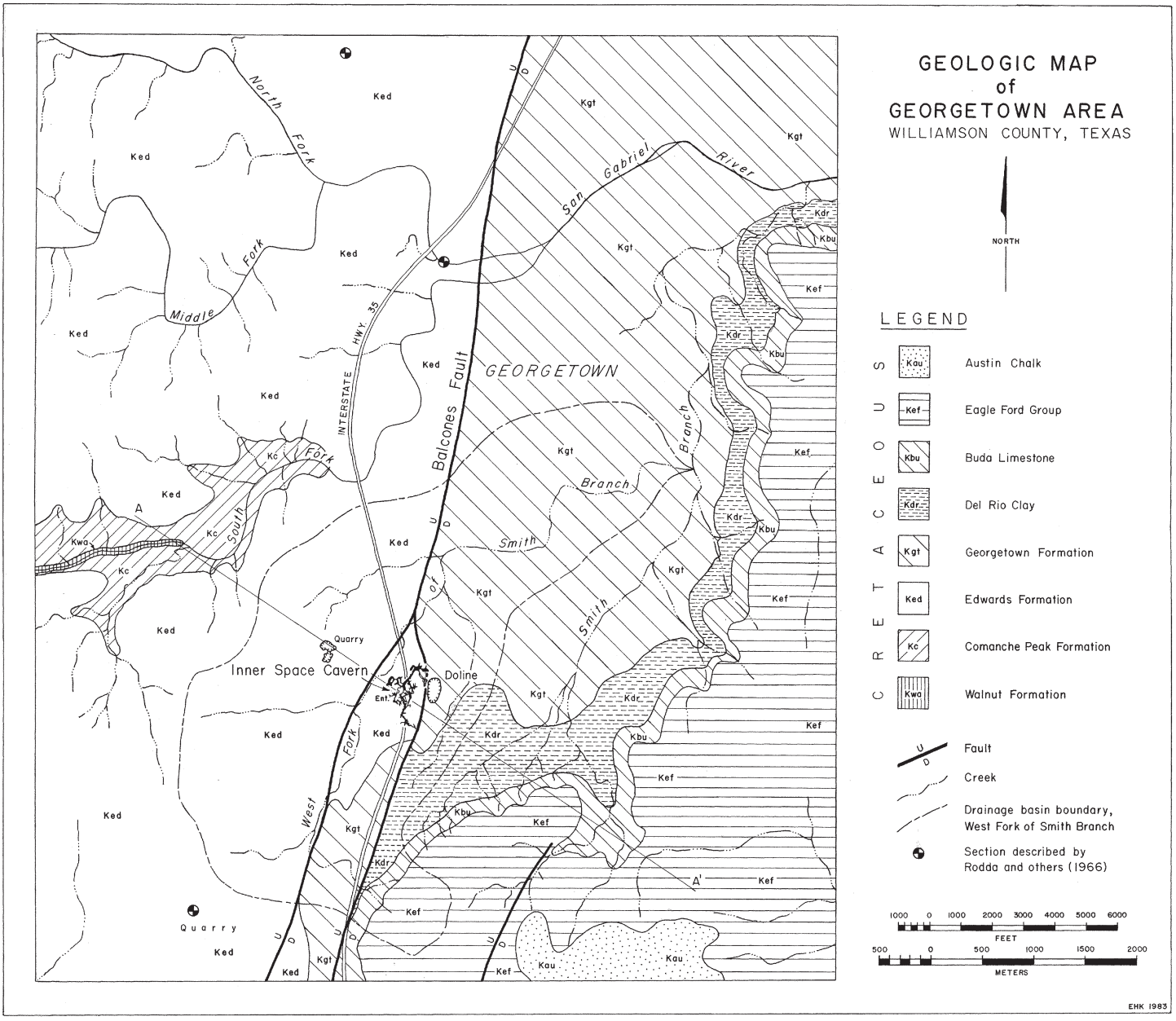


Figure 150: Geologic map of the Georgetown area, Williamson County, also showing drainage, and Inner Space Cavern. Geologic section A-A' is shown in Figure 151. Geology modified from maps of Ward (1950), Walls (1950), and Barnes (1974a).

including the 6 of more than 80 documented caves in the county known to exceed 300 m in length (Reddell and Finch, 1963). Caves of the area generally consist of wide (4 to 15 m) horizontal corridors that are confined to particular strata (see maps in Reddell and Finch, 1963, and Byrd, 1975). Several major caves have sizeable side-passages (e.g. The Bat Well, Beck Ranch Cave, Chinaberry Cave, Coffin Cave); others are large, singular conduits (e.g. Cobb Caverns) or have extensive, maze-like sections (e.g. Inner Space Cavern, Steam Cave).

Few caves were recorded in the Georgetown area prior to 1960 (Vinther, 1948; Widener, 1958). Cobb Caverns, 6.4 km north of the area, was opened commercially for a brief period beginning in 1962 (Reddell and Finch, 1963; Reddell, 1964b). However, Inner Space Cavern, the most significant cave in the area (more than 6 times longer than any other) and one of the longest in the state, was discovered in 1963 (Figure 152). At that time a crew from the Texas Highway Department drilled several coreholes to examine and test the foundation for a proposed overpass for Interstate Highway 35. At least four holes intersected cavities in the subsurface. Subsequently, one hole was enlarged to admit investigators from the Highway Department who quickly determined that a sizeable cave existed beneath the highway (Lindsley, 1963b; Kunath, 1965; Lehnhardt, 1970; Elliott, 1970b; Jasek, 1975c).

The Texas Speleological Association explored and mapped nearly 2 km of passages in the new cave from November 1963 to early 1964 (Lindsley, 1963b; anonymous, 1963b; Lehnhardt, 1970; Russell, 1975a). Originally called Laubach Cave, the well-decorated cave was developed and opened commercially as Inner Space Cavern in July, 1966 (Pinkard, 1966; Todd, 1966; Elliott, 1970b). Access thereafter has been through a tunnel excavated into the east side of the cave (Figure 152). Exploration, mapping, and study have continued sporadically since that time, with discovery of new passages (Jasek, 1975c; Russell, 1975a, 1980) and investigations of the cave's geology, internal environment, biota, and paleontological remains (Lundelius, 1975a,b).

Gravimetric traverses over the surface south of the new entrance (performed by A. Richard Smith and John Fish during the 1960's) indicate that at least two unentered passages may exist in that vicinity (Smith, 1966; Lehnhardt, 1970; Elliott, 1970b; Russell, 1975a, 1980). During the early visits explorers noted the variety of speleothems and mineral deposits. Reddell and Finch (1963) and Tart (1964) commented on sulfide minerals and calcitic boxwork in the northern part of the cave. Harmon (1969, 1970a,b) investigated the geochemistry of seeping water associated with active precipitation of calcite at the Flowing Stone of Time in the Outer Cathedral (Figure 152). He determined that precipitation of calcite from solution owing to loss of dissolved carbon dioxide to the atmosphere of the cave exceeds that resulting purely from evaporation by an order of magnitude. Speleothems taken from the cave were used in an investigation to verify that isotopic analysis of modern speleothems provides temperatures equivalent to those presently observed in the caves (Schwarcz and others, 1976). This method, which involves determination of deuterium/hydrogen ratios in fluid inclusions and $^{18}O/^{16}O$ ratios in calcite of speleothems and the calculation of paleotemperature, has been subsequently used in North America and elsewhere to help investigate paleoclimates (e.g. Harmon, 1975; Harmon and others, 1975, 1978).

Perhaps one of the most significant scientific finds in caves of the Edwards Plateau has been the discovery of considerable fossil remains of vertebrate fossils in Inner Space Cavern (Reddell and Finch, 1963; anonymous, 1963b; Frank, 1964; Slaughter, 1964, 1966a; Choate and Hall, 1967; Lehnhardt, 1970; Elliott, 1970b; Lundelius and Slaughter, 1971; Lundelius, 1975a,b). An annotated faunal list of fossils obtained from deposits in this cave has been compiled by Elliott (1970b, p. 204, 207).

Boyer (1979) investigated debris from talus-cones of Bone Sinks 2 and 3 in Inner Space Cavern which are known to contain remains of Pleistocene vertebrates. His analysis of clay minerals, in comparison with surficial and vugular deposits, indicated that the sediments may be relict soils from the late Pleistocene.

Few hydrogeologic investigations have been carried out in Williamson County. Cumley and others (1942) and Follett (1956e) have measured water-levels of wells and springs in the Georgetown vicinity. Cronin and others (1963, p. 70-73) present an overview of the water-bearing properties of the Edwards Formation, and Brune (1975) briefly describes several large springs in Williamson County. Water-quality and problems of land-use in carbonate terranes of the county have recently come under investigation (Hunt, 1973; Evans, 1974; Plamandon, 1975), indicating that an understanding of the hydrogeology of the local Edwards karstic aquifer is becoming

DEVELOPMENT OF CAVES

I have chosen Inner Space Cavern as the focus of study for the Georgetown area. The cave exhibits characteristics that are typical of most caves along the outcrop of the Edwards Formation and Balcones Fault Zone in Williamson County: (1) Inner Space Cavern has developed along a favorable stratigraphic horizon,

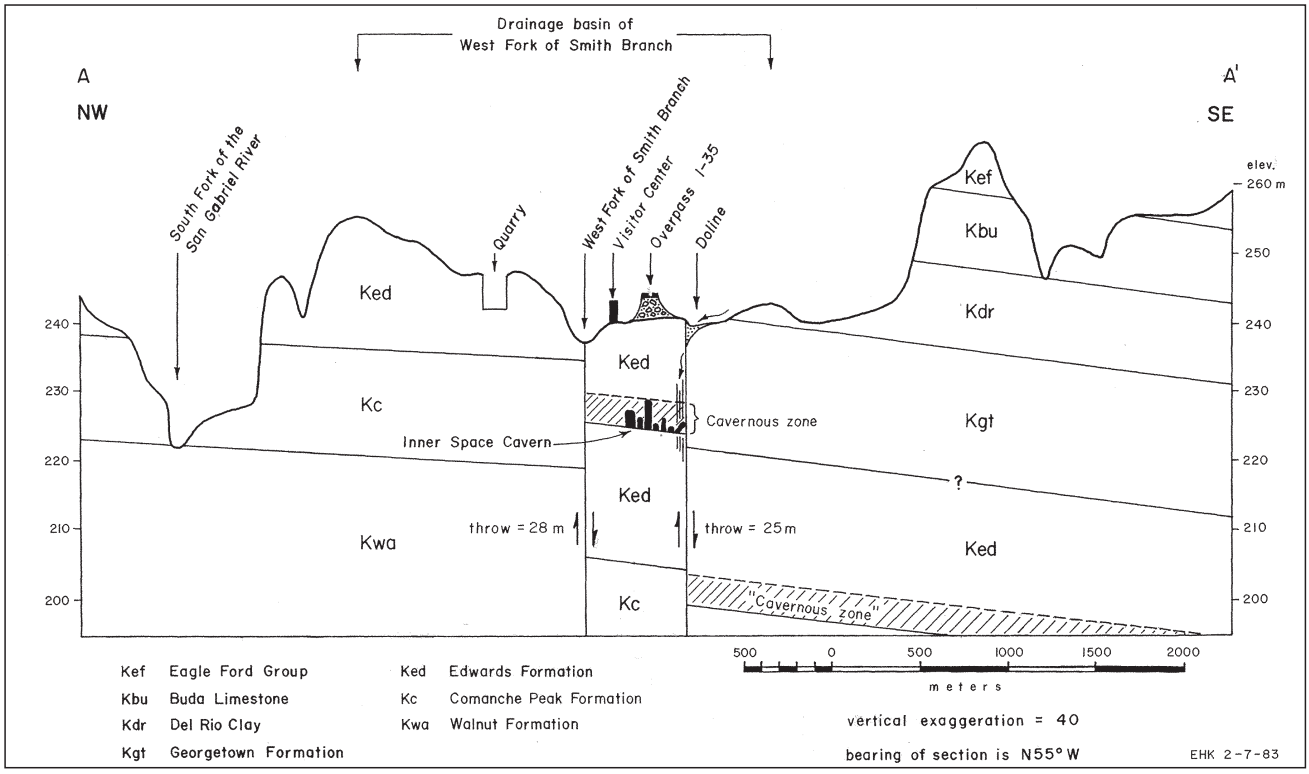


Figure 151: Geologic section A-A' through the Georgetown area, showing position of the cavernous zone of the Edwards Formation and Inner Space Cavern. For location, see Figure 150. Based on data in Ward (1950) and Barnes (1974a).

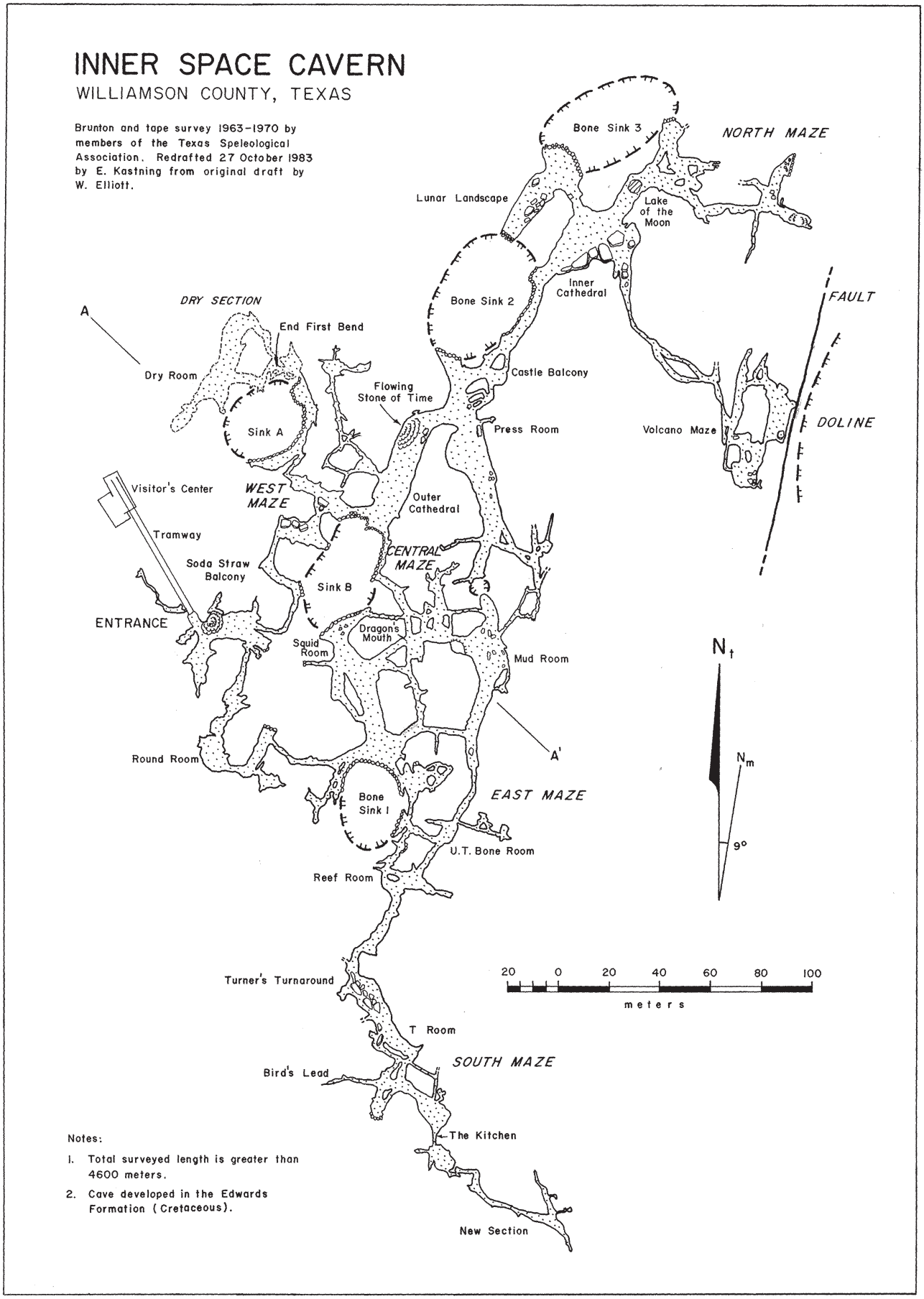


Figure 152: Map of Inner Space Cavern. Modified from Elliott (1970b, 1975) with geologic features added.

(2) it consists of main conduits of sizeable cross-section, tributary passages, and sections of mazes, and (3) linear segments of passages suggest structural control, particularly from regional faulting.

Description of Inner Space Cavern

The entrance to Inner Space Cavern is 2.9 km south-southwest of the center of the city of Georgetown and 180 m west of Interstate 35, approximately 2.1 km south of where it crosses the South Fork of the San Gabriel River (Figure 150). The cave has a mapped length of more than 4.6 km, an end-to-end extent of 644 m, and internal relief of approximately 25 m (Figure 52). Currently the cave is one of the longest mapped caves in Texas (Appendix C), and it is the longest cave of the northeastern part of the Balcones Fault Zone.

There are no natural entrances to the cave that have been open during historic times. Two artificial entrances provide access to the cave. The 13-m deep vertical borehole through which the cave was originally explored pierces the ceiling of the Inner Cathedral in the northern part of the cave. A 52 m-long included excavated tunnel enters the west-central part of the cave near the Soda Straw Balcony and presently serves and the entrance and exit for tours.

Inner Space Cavern is a complex maze-like cave in which both network and branchwork patterns are present (Figure 152; see Palmer [1975] for types of maze patterns). The overall maze-like character is produced through the integration of three primary types of passages: (1) trunk-passages, (2) linear segments of passages, and (3) collapsed sections (up to 20 m wide and 10 m high), and form the "backbone" of the cave. Linear segments of passages are generally shorter, linear in plan-view, small in cross-sectional area, and form angulate maze-patterns situated adjacent to trunk-passages. Collapsed passages are typically curved in plan-view, with one wall composed of collapsed material They commonly skirt the periphery of debris-cones within collapsed dolines and, in places, are blocked by breakdown. Examples of passages of each type are given in Table 11.

The larger passages form a somewhat dendritic pattern in which convergence occurs in a northerly direction (Figure 152). This is most pronounce in the trunk-passages; however, numerous linear segments of passages and some collapsed passages resemble "tributaries" to the system as well.

Most of Inner Space Cavern has developed along a single level positioned approximately 13 to 20 m below the surface. Strata of the Edwards Formation along this horizon are typically high (85 to 95 percent) in content of calcite (Rodda and others, 1966, p. 269-275) and exhibit moderate to honeycomb porosity. Some beds of the Edwards, both subjacent and superjacent to the horizon of the cave, are locally dolomitie and vary between 55 and 85 percent in content of calcite.

The northern half of the cave is relatively dry, with no perennial streams; however, the southern half of the cave often floods. During wet periods water inundates the trunk-passage extending from the Mud Room to the Chapel, tampering exploration for up to four months in some years (Elliott, 1970b; Russell, 1975a). This section is a few meters lower in elevation than the rest of the cave and ponds during wet seasons. The Volcano Maze also floods occasionally.

Locally, vertical seepage of surficial water into the cave is very pronounced. Deposits of dripstone and flowstone are profuse in some sections and several speleothems are massive (e.g. The Flowing Stone of Time in the Outer Cathedral; see Harmon [1969. 1970b] for description). Soda-straw stalactites are ubiquitous and some in the New Section southeast of the Kitchen are nearly 2 m long (Lehnhardt, 1970; Elliott, 1970b). Soda straws 0.5 to 1.0 m in length are relatively common throughout the cave (e.g. in the Soda Straw Balcony near the entrance). Helictites are also prevalent in many passages and soon attain lengths of 20 cm (Reddell and Finch, 1963, p. 55).

Table 11: Types of Passages in Inner Space Cavern[a]

Trunk Passages	Linear segments of passages	Collapsed Passages
Lake of the Moon	North Maze	Passages at Bone Sink 3
Lunar Landscape	Volcano Maze	Passages at Bone Sink 2
Inner Cathedral	West Maze	Passages at Sink A
Press Room	Central Maze	Passages at Sink B
Dry Room	East Maze	Passages at Bone Sink 1
Outer Cathedral	South Maze	
Squid Room		
Mud Room		
U. T. Bone Room		
Reef Room		
Turner's Turnaround		
T Room		

[a]Refer to Figure 152 for locations.

The cave is floored by a variety of sediments. Relatively thick deposits of fill of clay and silt occur in sections that flood. The passage between the Press Room and Mud Room has been completely filled at one point, blocking both exploration and flow of water. Other sections of the cave are floored with silt-to-sand-sized particles, and slabs of breakdown litter floors in the vicinity of collapsed dolines.

Considerable vertebrate fossils have been discovered in the debris-cones of the collapsed dolines and beneath sediments in the trunk-passages proximal to the doline-deposits. Many bones have been retrieved, dated by the C-14 method, and determined to be from mammals of the late Pleistocene (Choate and Hall, 1967; Frank, 1964; Lundelius, 1975a,b; Lundelius and Slaughter, 1971; Slaughter, 1964, 1966a). These finds suggest that the cave was open at various times during the Pleistocene, approximately 13 to 45 thousand years ago (Slaughter, 1966a; Elliott, 1970b; Lundelius, 1975a,b).

Deposits of sulfide minerals have been reported from the northern end of the cave, in the vicinity of the Inner Cathedral and Lake of the Moon. Reddell and Finch (1963) report crystals of galena, up to 5 m in size, along solutionally enlarged joints. Tart (1964) describes crystals of pyrite of similar size from the same locality and notes that many exposed crystals have weathered to limonite. Sulfide deposits are associated with crystals of calcite and calcitic boxwork similar in form to that found in Wind Cave, South Dakota, and described by White and Deike (1962) and Palmer (1981b).

Inner Space Cavern lies near the center of the drainage basin of the West Fork of Smith Branch where the surface of the land slopes gently (3.7 m/km) to the northeast. Fluvial dissection is moderate in the vicinity of the cave but increases northwardly to the San Gabriel River. Passages of the cave lie approximately 22 m above local baselevel at the San Gabriel River.

Further descriptive and historical information about the cave may be found in Lindsley (1963b), anonymous (1963b), Reddell and Finch (1963), Reddell (1964b), Pinkard (1966), Todd (1966), Lehnhardt (1970), Elliott (1970b), Jasek (21975c,d), Fieseler and others (1978), and Russell (1975a, 1980). Photographs of many features of the cave are included in Pinkard (1966), Todd (1966), Fieseler and others (1978), and in various issues of the *Texas Caver* (1963 to the present).

Speleogenesis

Superposition of three distinct types of passages has given Inner Space Cavern an inherent complexity that would not normally be expected in an area where stratigraphic, structural, and topographic settings are rather uniform and uncomplicated. This suggests that development of the cave has responded to some abrupt changes in the hydrogeologic environment, wherein more recent morphogenetic characteristics have been overprinted in older ones.

Lithologic Control

Inner Space Cavern has developed along a particular horizon that lies approximately in the middle of the Edwards Formation (Figure 151). The strata of this zone consist of medium- to thick-bedded, highly pure limestone with few chert beds. Most of the rock is aphanitic, crystalline, and honeycombed and some beds contain abundant molluscan fossils indicative of a biostromatal or reefal environment. These characteristics have given this horizon a considerable primary porosity which has apparently augmented later flow of groundwater under phreatic conditions.

Many passages in the cave have shelves or re-entrants along the wells where excavation has followed bedding-plane partings. These may represent burrowed zones in the original stratigraphic sequence that were later rendered vugular during diagenetic and dissolutional modification. In fact, most maze-like passages of low ceiling-height, which are products of ponding and flooding (see below), follow bedding-plane partings. In general, passages are conformable with the strata. Those to the west and slightly updip are slightly higher than those along the eastern side of the fault block (Figure 151).

The argillaceous Georgetown Formation has largely precluded excavation of caves east of the major faults at Inner Space Cavern. However, a large doline 460 m east of the entrance, and just east of the eastern-most major fault, has developed within the Georgetown Formation (Figures 150, 151, 152). Here, some of the runoff from the relatively impermeable uplands to the south has been diverted through

fractures to the cave-system in the vicinity of the Volcano Maze. Dissolution and sapping in this area of infiltration have produced the doline adjacent to the fault-plane (Figure 151).

Structural Control

A preponderance of linear segments of passages on the map of Inner Space Cavern (Figure 152) suggests that development of the cave has been largely guided by structure. This is confirmed by comparison of trends of passages with the attitude of bedding and regional patterns of fractures.

Relationship to Attitude of Beds. Beds within the fault-block containing Inner Space Cavern dip about 0.2° to the southeast. Trunk-passages are typically aligned parallel or subparallel to the strike of the bedrock. In places, however, short segments are oriented otherwise, including parallel to the direction of dip. Horizontal development of trunk-passages along the strike and within relatively soluble strata is indicative of excavation in the shallow-phreatic zone where the piezometric surface is nearly horizontal. Inspection of the morphology of passages supports this interpretation. Trunk-passages are typically elliptical in cross-section where they have not been modified by collapse. Ceilings of trunk-passages are essentially horizontal except in areas of extensive modification by floodwater (see below). Where gradients of floors can be inferred, it is seen that they too are nearly horizontal.

Cross-sectional areas of trunk-passages are slightly larger in the northern part of the cave, particularly where two or more passages merge in this direction (Figure 153A). Northwardly confluent conduits and the northward surficial drainage of the area indicate that movement of groundwater within the evolving cave was to the north.

Trunk- and tributary-passages along the western flank of the cave are 0.5 to 3.0 m higher in elevation than those to the east. These locations correspond to equivalent stratigraphic positions within the cavernous horizon (Figure 151), a situation that does not preclude simultaneous development of the passages. For Inner Space Cavern this obviates the argument that passages of lower elevations are necessarily younger (owing to a progressive lowering of the piezometric surface, for example).

On the other hand, a few tributary-passages of high and narrow cross-section along the western flank of the cave are slightly higher in elevation than are the trunk-passages with which they communicate. This is true of passages in the North, Volcano, and South Mazes (Figure 152). Sections of these passages are discordant with strata, particularly avenues that trend across the strike of the bedrock. These passages have apparently served as collectors of water that has infiltrated through dolines and fractures in proximity to the large faults along the eastern margin of the cave. In this capacity these conduits may have evolved in the vadose zone as dip-tubes. The passage connecting the Volcano Maze with the main cave is canyon-like, appearing as though its floor may have become entrenched and graded by free-surface streamflow.

Inclined, vadose, canyon-like passages aligned down the dip and horizontal, phreatic, tubular passages trending along the strike are common in gently dipping strata (Palmer, 1977, 1981a). The relationship of these two types of passages to one another in Inner Space Cavern is similar to that described in Longhorn Cavern, Burnet County (Chapter 9).

Relationship to Fractures. Nearly all passages in Inner Space Cavern consist of linear segments (Figures 152). Internal examination of passages shows that vertically oriented fractures are observable in ceilings and walls throughout the cave, and many are positioned longitudinally along passages. Horizontal extents of fractures are variable; most range between 2 and 90 m in length. Where vertical extents are observable, joints appear to extend completely through the cavernous zone; hence, lithologic differences in the rock have not significantly influenced fracturing within that horizon.

Orientations of segments of passages are shown in the rose-diagram of Figure 154. Because of the strong correspondence of passages with fractures, the rose-diagram can be viewed as a summary of the orientation of fractures. There are distinct concentrations of fractures (segments of passages) in the intervals N20° W to N 40 E and N 50°-90° W. However, closer examination shows there are two primary sets of fractures at N 10°-20° E (set I) and N 80°-90° W (set II) and corresponding conjugate sets at N 30°-40° E (set IIIA) and N 10°-20° W (set IIIB). A comparison of these orientations with those of major faults (Figure 150) shows that most segments of passages along fractures are generally either parallel or orthogonal to the faults, whereas most others are parallel to the conjugate directions.

Patterns of fractures are consistent with the tectonic history of the region (Figure 155). Balcones faulting during the early Miocene was primarily extensional in the

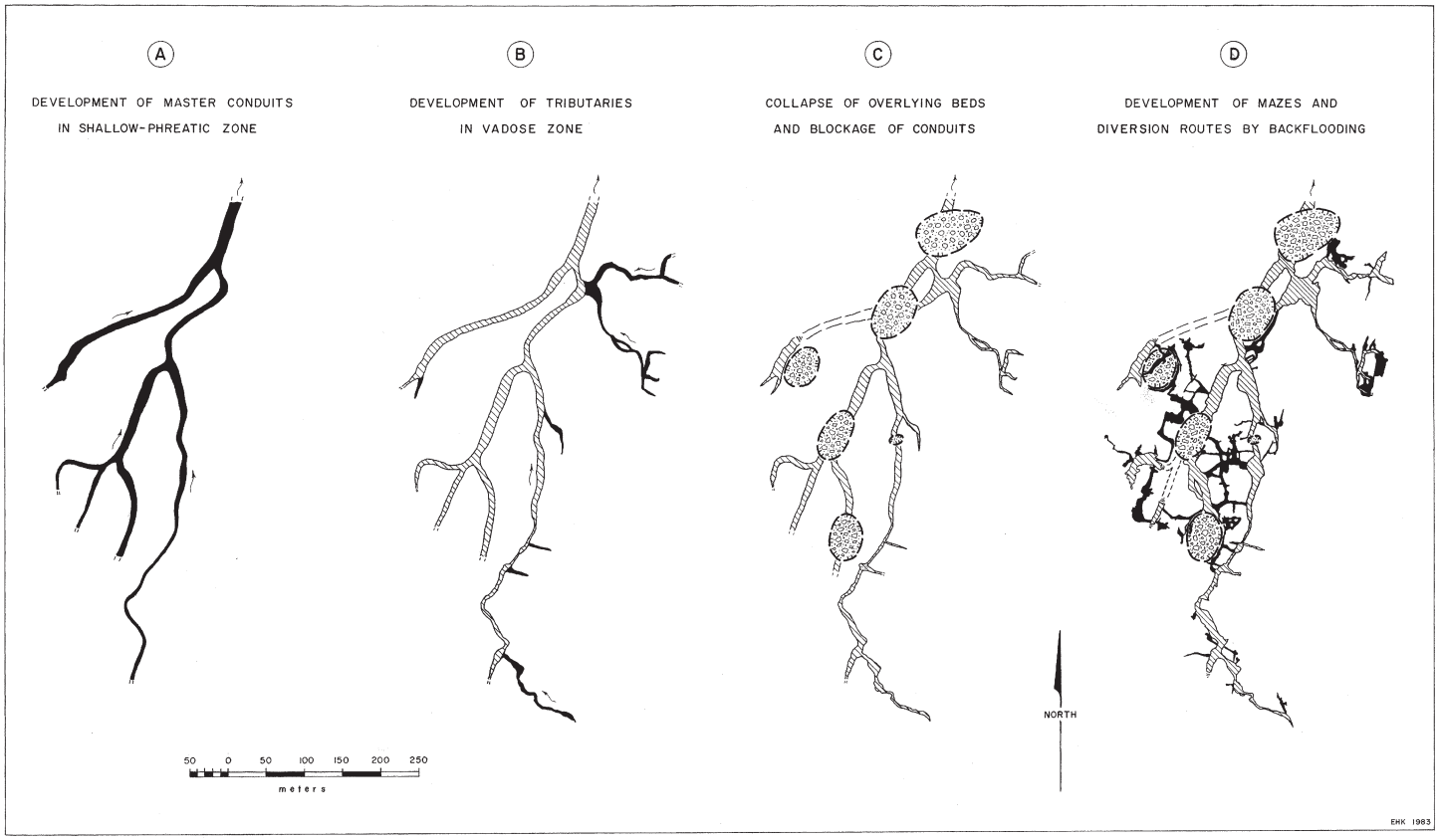


Figure 153: Origin of Inner Space Cavern. (A) Development of master conduits (trunk-passages). (B) Growth and integration of tributary-passages and vadose feeders. (C) Collapse of dolines into larger passages in the vicinity of intense fracturing. (D) Modification through periodic flooding and development of extensive maze-like networks.

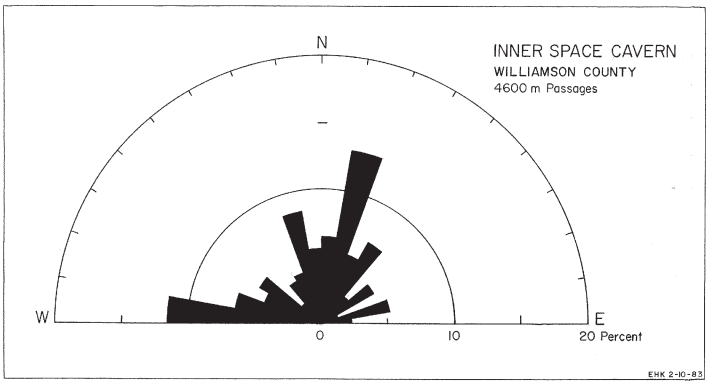


Figure 154: Rose-diagram showing orientations of linear segments of passages in Inner Space Cavern. With few exceptions segments have been excavated along fractures. Compare with Figure 155.

northeastern Edwards Plateau; simultaneous uplift in the plateau to the west and subsidence in the Blacklands to the east created a series of *en echelon* step-faults (Foley, 1926; Jones, 1928; Link, 1929; Sellards, 1934; Perry, 1940: weeks, 1941, 1945b; Fowler, 1956, 1957). Orthogonal and conjugate sets of fractures such as those observed here are typical in rocks deformed under simple extension. Because tensionally produced joints tend to be fairly open (in comparison with those produced under compression) and thus augment circulation of groundwater, it is not surprising that nearly all avenues of excavation in Inner Space Cavern lie along them.

Small, subsidiary faults occur in various sections of the cave (Reddell and Finch, 1963). For the most part, they are of small displacement (0.5 m or less) and have merely behaved as open joints during speleogenesis. As such, small faults have operated "positively" in enlargement of conduits (Kastning, 1977c). A case in point is the Volcano Maze where small faults, striking parallel to the main Balcones fault and within 50 m of it, have guided development of at least two north-south trending passages (Figure 152). Seepage from the large doline created by solution-subsidence in the Georgetown Formation to the east may have moved downward along these faults during dissolutional excavation (Figure 151).

Elsewhere in the cave faulting has influenced excavation in a "negative" manner (Kastning, 1977c). The two large faults bordering the cave to the east and west have throws of approximately 25 m and 28 m respectively, bringing stratigraphic horizons favorable for development of caverns against others relatively resistant to excavation (Figure 151). Despite any openness in the faults incurred under tension, they behave as barriers to flow of groundwater normal to them. This is especially true along the eastern fault, where the Georgetown Formation is juxtaposed against the cavernous zone of the Edwards Formation. The overall effect is that groundwater is contained within the fault-block and restricted form moving eastwardly down the dip. Such compartmentalization is similar to that observed in the Bend, Burnet, New Braunfels, and Austin-San Marcos areas (Chapters 8, 9, 10, and 11 respectively).

Modification by Collapse. Passages of Inner Space Cavern are nowhere more than 25 m below the surface. Additionally, strata overlying the cave are thin- to medium-bedded, brittle, and intensely fractured. These conditions have made the cave highly susceptible to collapse. Ceilings have collapsed where trunk-passages have become too broad to support overlying expanses of limestone. In some places (e.g. the Inner Cathedral, Volcano Maze, and Turner's Turnaround) blocks and slabs of breakdown litter the floor, and cross-sections have been modified from elliptical to rectangular or trapezoidal. Elsewhere, especially where two or more trunk-or tributary-passages are close together or intersect, failure of the roof has been extensive, with some zones of collapse on the order of 40 to 70 m in diameter (Figure 152).

Faults and joints have significantly augmented collapse. Examination of the cave map (Figure 152) shows that major collapse dolines (Bone Sinks 1, 2, and 3; Sinks A and B) have formed in zones where fractures intersect. (In this visualization, extensions of linear segments of passages into collapsed areas are assumed to represent extensions of fractures.). Upward stoping of voids during collapse progressed to completion in some dolines, periodically opening the cave to the surface. Fossils of vertebrates from the late Pleistocene and Holocene found along walls of debris-cones and in sediments on the floor of the cave indicate that various entrances were open 13 to 45 thousand years ago (Slaughter, 1966a; Elliott, 1970b; Lundelius, 1975a,b). Difference in ages of debris-cones and entrances is suggested by differences in sediments within the cones an din the degree of cementation of these deposits (Lundelius, 1975b). Subsequent collapse, which may still be occurring today, has resealed these former entrances.

Collapse was aided by surficial processes as well as by subsurficial excavation. Denudation of the Edwards Formation and overlying strata progressed since the time of faulting, in response to changes in relief imposed by tectonism (see Chapter 2). Moreover, downward sapping of weathered limestone through open fractures, particularly in zones of high density of fractures, led to the development of dolines formed by solution-subsidence over the subsurficial cavernous zones. Concomitant solutional thinning of the overlying limestone from both above and below accelerated collapse.

Hydrogeological and Topographic Controls

Horizontal passage and their confinement along strike and within a favorable stratigraphic horizon imply that excavation occurred just below the piezometric surface. During enlargement of master conduits, groundwater flowed to the north-northeast, along a gentle hydraulic gradient roughly parallel to the evolving surface of the drainage basin of the West Fork of Smith Branch. Phreatic flow produced at least three major conduits which were confluent to the north (Figure 153A).

Discharge at baselevel probably occurred along the South Fork of the San Gabriel River at or near the point where it is crossed by the Balcones Fault (Figure (150). Although not established during this study, the conjectured northward route of the flow of groundwater lies adjacent to and west of the fault and largely follows fractures parallel to the trend of the fault and largely follows fractures parallel to the trend of the fault. Discharge along faults has been documented in the area. Brune (1975, p. 84) notes that Manske Branch Springs, 9.7 km east of Georgetown, flows at 65 liters per second.

Groundwater from the fault-block containing Inner Space Cavern presumably flowed northwardly within the Edwards Formation, although perhaps on a horizon lower than the cavernous zone in the vicinity of the cave. Today, lower beds probably conduct water northwardly from the locality of the cave.

The elevation of the piezometric surface at the cave remained in position long enough for master conduits to attain sufficient size and integration (Figure 153A). Originally water may have flowed along two or more parallel avenues. Upon solutional enlargement, hydraulic gradients between adjacent channels may have become steep in comparison to those to the north along strike. As a consequence, smaller channels may have been pirated and diverted along cross-fractures, integrating the system into a dendritic network (Figure 153B; see Palmer, 1975, p. 57-58, for an explanation of development of branchwork-patterns).

Piezometric levels are generally tied to incision by streams and to local baselevels. Waterlevels at Inner Space Cavern dropped in response to downward cutting of the San Gabriel River and denudation within the drainage basin of the West Fork of Smith Branch. Eventually the waterlevels dropped beneath the level

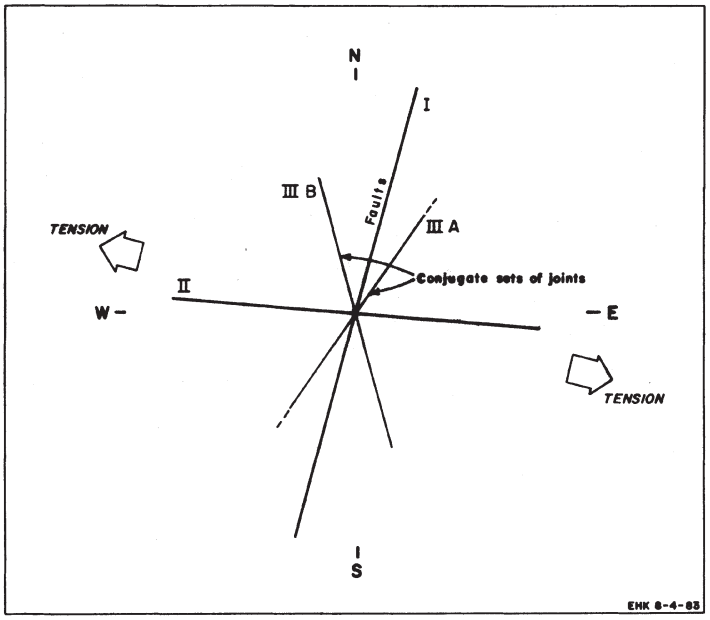


Figure 155: Schematic illustration of extensional tectonism in the Georgetown area. Alignments of resultant fractures are shown in relation to deformational stress. Compare with Figure 154.

of phreatically produced conduits. Based on data from wells published in Cumley and others (1942) and Follett (1956a), the present-day piezometric surface lies between 10 to 30 m below the level of the cave. In the ensuing vadose phase the cave became modified by collapse of some ceiling-beds, development of vadose feeder-passages carrying surface-derived recharge, periodic flooding, and deposition of sediments and speleothems.

Collapse of ceiling-beds was probably enhanced by removal of buoyant support. This initiated the stoping phenomena that would ultimately produce the large, collapsed dolines encountered in the cave (Figure 153C).

Surficial water entering open joints in the vicinity of the Balcones Fault easily percolated downward through the vadose zone to the piezometric surface. Some of this recharge was concentrated by intersecting systems of joints (Figure 153B) and transmitted to the trunk-passages at times when the latter were either still inundated or after they were drained. The best examples of vadose drainage are the easternmost passages of the cave which have collected water from the area of the fault and have transmitted it westwardly to the trunk-passages. Relatively high and narrow canyons along such routes typify vadose conditions, where floors of passages are lowered and incised by freely flowing streams.

Trunk-passages were severed by large-scale collapse of dolines (Figure 153C). Because the piezometric surface was still relatively high, the cave was subjected to flooding during wet seasons. Flooding may also habve been enhanced by wetter and colder climates during the Pleistocene, at a time when many other caves on the Edwards Plateau are known to have flooded. Flood-pulses entered the cave from time to time, completely inundating all passages. Turbulent flood-waters, undersaturated with respect to calcite and under locally steep, hydraulic gradients, are known to be highly aggressive in the excavation of caves (Malott, 1937; Thrailkill, 1964, 1968; Palmer, 1971, 1972, 1975). In Inner Space Cavern, floodwaters impounded behind blockages at collapsed dolines rapidly excavated nearly fractures and produced numerous joint-spurs (short, dead-end side-passages along joints), enlargement of passages upstream from blockages, and routes of diversion around collapsed areas (Figures 152, 153D).

Flooding and vadose recharge may have transported much of the elastic sediment which now partially fills most conduits. Some fill is rich in vertebrate, skeletal material, indicating that its deposition may date from the late Pleistocene. This is supported to some extent by analyses of clay minerals from sediments of bone-deposits, which suggest that the sediments may, in part, be relict soils washed into the cave (Boyer, 1979).

Deposition of speleothems is actively occurring throughout the cave (Harmon, 1969, 1970a,b). The ubiquity of deposits of calcite, in particular rapidly forming soda-straw stalactites, and the steady availability of seepage suggest that infiltration is presently occurring along most fractures. Openness of fractures and an overall shallow depth of the cave undoubtedly contribute to high rates of percolation.

Soil in the vicinity of the cave may provide some explanation for present-day conditions of infiltration and deposition of speleothems. The Crawford stony clay and Denton stony-clay soil, which overlie the outcrop of the Edwards Formation within the fault-block and to the west, are typically shallow (5 to 90 cm) nonarable soils containing fragments of limestone and chert (Templin and others, 1938). The deeply developed, clay-rich zone over the Georgetown significantly impedes infiltration. In contrast, surficial runoff over the Edwards can more easily penetrate the shallow, stony soils overlying this fractured limestone.

High rates of infiltration through the Edwards and its shallow mantle of soil may explain the wet character and profusion of speleothems in Inner Space Cavern. However, some water in the soil over the Edwards Formation may have been transmitted to the west from near-surficial, clayey soils overlying the Georgetown Formation. Percolation through deep. Organically rich, clayey soils will increase the content of carbon dioxide in such waters, making them more aggressive in dissolving limestone in the vicinity of the cave. Because these waters are then able to hold more dissolved calcite, they contribute more material for deposition of speleothems as seepage enters the atmosphere of the cave, where partial pressures of carbon dioxide are substantially reduced.

Chronology of Speleogenesis

The following chronological sequence is proposed for the origin of Inner Space Cavern:

1. Rocks of the Fredericksburg and Washita Divisions, including the Edwards, Georgetown, and later formations, were deposited during the Lower Cretaceous, and varied considerably in lithic character. During diagenesis, the Edwards Formation became a hard, brittle, highly porous, relatively pure limestone, in contrast to most of the overlying sediments which were considerably softer, less porous, marly, and dolomitic.

2. Rocks in the Georgetown area were faulted in the early Miocene. Tensional forces produced step-faults and fault-blocks striking generally north and a fault-scarp approximately 50 m in height in the present vicinity of Inner Space Cavern. Bedrock was extensively fractured during this time, forming joints of two primary orthogonal sets and two conjugate sets (Figures 154 and 155).

3. Erosion of the uplands west of the Balcones Fault Zone began in the late Tertiary and continued through the Pleistocene. During this interval the surface over Inner Space Cavern was gradually lowered in response to incision of the San Gabriel River and its tributaries. Hydraulic gradients in the subsurface steepened toward the river

4. As northward drainage evolved on the surface, so did northward paths of flow of groundwater. The piezometric surface within the Edwards Formation remained positioned above the present, cavernous zone for sufficient time to allow development of trunk-passages under shallow-phreatic flow (Figure 153A).

5. Infiltration entering carbonate units within the fault-block, near the upstream reaches within the drainage basin of the West Fork of Smith Branch, was transmitted northward through available fractures. This led to the development of a dendritic network of conduits consisting of master conduits (trunk-passages) and tributaries (Figure 153B). Throughout its evolution the system of flow remained confined between the two faults of large displacement along the east and west boundaries of the fault-block. It is presumed that water leaving the northern end of the block was transmitted northward to the San Gabriel River, along fractures adjacent to the Balcones Fault and within the Edwards Limestone (Figure 150).

6. The piezometric surface gradually lowered during continual erosion of the surface, eventually dropping beneath the cavernous zone. Some minor collapse of ceiling-beds may have occurred at this time. Infiltration from the surface, entering along joints and faults in the vadose zone, became integrated along tributaries leading to the older cave (Figure 153B). Many of these avenues are canyon-like in cross-section, a characteristic of vadose development.

7. Further denudation of the landscape, coupled with stoping within larger passages, led to large-scale collapse (Figure 153C). This was augmented by the abundance of fractures and the close spacing of trunk- and tributary-passages.

8. Inner Space Cavern apparently underwent periodic flooding since (and perhaps during) the time of collapse. Flooding may have been induced by colder and wetter pluvial climates during the Pleistocene. Vadose waters coursing through segments of trunk-passages severed by collapse were impounded behind terminations of debris. Backflooding produced enlargements of passages upstream of blockages, excavations of spur-passages along joints, and propagations of diversion routes around zones of collapse. Joint-spurs and routes of diversion along fractures have made the cave maze-like (Figures 152, 153D).

9. Slumping of debris continued along the sides of collapsed dolines. At times entrances were opened to the surface, providing a means for animals and their remains to enter the cave. Sediments were deposited in the cave by three mechanisms: (1) washing through entrances in dolines, (2) transmission along open joints through solution--subsidence and sapping of dolines, and (3) transportation by surges of floodwater. In some cases these sediments form the matrix of deposits of vertebrate fossils. Deposition of speleothems began at about this time.

10. Further collapse eventually sealed all entrances. Deposition of calcite and small-scale flooding continued into the Holocene Epoch.

SUMMARY

Development of caves in the Georgetown area has been largely confined to relatively porous and soluble beds within the highly fractured Edwards Formation. Large and complex caves have evolved in areas of intense faulting and jointing. Initially, caves developed as branching networks of conduits and consisted of tributary-channels which fed master-tubes (trunk-passages) that, in turn, conveyed water along strike under shallow-phreatic flow. Where fault-blocks were bordered by faults of large displacement, water had been guided to springs at base level along the axes of fault-blocks. Surficial degradation caused ceilings of large passages to collapse where they had been weakened by fractures or by enlargement of passages. Some major conduits had been blocked by collapse, and water ponded and flooded large sections of some caves, ramifying existing passages into extensive maze-like networks. The multigenetic development of caves in the Georgetown area is best exemplified by Inner Space Cavern. This mode of origin and extensive deposits of bones of Pleistocene animals and of speleothems make Inner Space Cavern one of the most geologically important caves of the Edwards Plateau.

13
GRANITIC PSEUDOKARST IN THE LLANO BASIN

INTRODUCTION

Karstlike landforms have developed on exposed, granitic rocks at several localities in the Llano Basin. Generally, these features are products of processes other than dissolution and can be properly classified as pseudokarst (for definitions of pseudokarst and problems in terminology, see Chapter 1, Appendix A, Otvos (1976), and Beck (1979, 1980))

Granitic pseudokarst is somewhat incidental to the main thrust of this study; however, a discussion of its characteristics and origin is included here for completeness and gives perspective to the different processes and landforms characteristic of weathering of soluble versus relatively insoluble rock. Moreover, analysis of these minor and isolated forms provides additional data on geomorphic processes active in central Texas following denudation of the Llano Basin during the Tertiary and Quaternary. Findings discussed in this chapter are based on studies performed at Enchanted Rock State Natural Area from 1974 to 1978 (Kastning, 1977a,b, 1979, 1980d)

Geographic Setting

Enchanted Rock State Natural Area is located in southwestern Llano County in the south-central part of the Enchanted Rock U.S. G.S. 7.5-minute quadrangle. Enchanted Rock itself is the largest dome in a chain of domes bounded on the southeast by Sandy Creek and on the northwest by the watershed of Walnut Spring Creek (Figure 156). Most of the immediate area is drained by Sandy Creek. Keener Branch of Crab-apple Creek drains much of the area southeast of Route 965. These streams are tributaries of the Colorado River.

Granitic hills rise from the gently rolling floors of the basins of the creeks, giving the area an overall hummocky appearance. The Enchanted Rock dome is oblong in plan-view, measuring 760 m by 520 m, and it rises 137 m above Sandy Creek to an elevation of 556 m above sealevel (Figures 156 and 157).

Previous Geomorphic Investigations

Enchanted Rock is a well-known physical landmark in the Llano Basin. It was noted as an impressive natural feature in early reports of geological reconnaissance in central Texas (Buckley, 1966, p. 13; Comstock, 1891). The vicinity of Enchanted Rock was first mapped in detail by Paige (1911, 1912a) and later by Barnes, Dawson, and Parkinson (1947), Hutchinson (1956, 1960), and Barnes (1981a).

Landforms and weathering at Enchanted Rock and nearby domes were first described by Blank (1951a,b). Shepherd (1975a,b, 1979) investigated the form and evolution of stream-channels of the Sandy Creek watershed and briefly described Enchanted Rock and nearby Watch Mountain. Begle (1978), Patton and Folk (1978), and Folk and Patton (1982) discussed the development of grus based on investigation of weathering of granite in the Llano Basin.

Enchanted Rock Cave near the summit of the dome has been known for some time. Trail markings leading to its entrance have been in place for nearly 30 years (Moss and Moss, 1956). Russell (1968) reported a length of 90 m for the cave. However, descriptions and maps of the cave were published for the first time only recently (Smith, 1974; Kastning, 1977a,b, 1979; Fieseler and others, 1978, p. 34; Twidale, 1982a, P. 281-82).

GEOLOGIC SETTING

Enchanted Rock is composed of porphyritic leucogranite forming the outer zone of the Enchanted Rock Batholith, a Precambrian intrusion of Town Mountain Granite (Figure 158). The rock consists of quartz, microcline, sodic andesine plagioclase, calcic to sodic oligoclase, and less than five percent biotite (Hutchinson, 1953, 1956, 1960). Southeast of Route 965 the bedrock is generally Packsaddle Schist, but wall-rock adjacent to the batholith is locally recrystallized. Aplite, aplogranite, or quartz-rich zones occur locally (Figure 158). An example of the latter is Turkey Rock, a castellated dome just east of Enchanted Rock. Quaternary alluvium blankets much of the valley of Sandy Creek, and the channel of the creek is choked with granitic gravel derived as detritus from weathering of nearby hills of granite and schist (Shepherd, 1975b, 1979).

Igneous rocks of the region are exposed as a large pear-shaped area covering 260 square kilometers in Llano and Gillespie County (Figure 158). Emplacement of the batholith into Valley Spring Gneiss and Packsaddle Schist has been dated at approximately 1046 plus or minus 12 million years ago (Leon Long, verbal communication, August, 1978). Keppel (1940) and Hutchinson (1953, 1956, 1960) have mapped structures exposed on the batholith. Fractures and flow-structures, such as lineations, flow-layers (schlieren), platy parallelism of phenocrysts, parallelism of crystals of feldspar, and alignment of xenoliths, indicate relative magnitudes and directions of stress during intrusion Some of these structures and the petrological variation within the batholith, as mapped by Hutchinson (1956, 1960), possibly reflect cells of flow during cooling and solidification of magma. Ultimately these relict structures may have influenced the origin and distribution of domes and lesser landforms developed on them.

The Llano region experienced successive periods of uplift from the Early Ordovician to the Late Pennsylvanian. Extensive surfaces of erosion covered by marine, sedimentary rocks indicate that the buried Precambrian surfaces had been elevated nearly 3000 m during this interval (Cheney and Goss, 1952). Subsequent erosion during the Triassic and Jurassic removed much of the Paleozoic rock. Shallow seas once more covered the region during the Cretaceous, during which time sequences of sandstone and limestone were deposited. Continental elevation perhaps at the end of the Cretaceous, but especially in the Miocene, brought the region well above sealevel once more, and erosion has since removed the Cretaceous rocks and additional Paleozoic rocks. Some exposures of Hickory Sandstone and Cap Mountain Limestone (Middle Cambrian) are still found masking parts of the batholith (Figure 158). Where uplift and erosion have been greatest, rocks of the Precambrian basement have been exhumed and subjected to erosion. This has resulted in a diversity of landforms, particularly on granitic rock.

Ancient patterns of drainage, developed on the Cretaceous rocks that once covered the Precambrian rocks, generally correspond to those presently known in the area owing to exhumation of Cretaceous topography. The Llano topographic basin evolved as the Cretaceous cover was stripped by gulfward drainage during the post-Miocene. At the present time the basin is primarily floored with Precambrian and Paleozoic rocks, and its boundary largely consists of erosional scarps developed on Cretaceous limestone. During the late stages in the evolution of the drainage of the Llano River (Kansan and post-Kansan), stream incised rapidly into pre-Cretaceous rocks and the limestone scarps receded. Some surficial streams on the limestone plateau surrounding much of the basin were captured by the drainage of the Llano River (Shepherd, 1975a,b, 1979). This in turn accelerated erosion of Precambrian and Paleozoic rocks in the period following deposition of Uvalde-type gravels.

WEATHERING AND EROSION OF GRANITE

Weathering of granite is a complex geochemical and geomorphologic process, and its mechanisms have been the subject of much debate. Granitic rocks are mineralogically simple, composed primarily of quartz, Microcline (orthoclase), plagioclase, biotite, and some hornblende. Accessory minerals occur in very small amounts and apparently have little effect on weathering.

The processes of weathering and erosion of granite may include to various degrees one or more of the following: (1) chemical alteration of minerals, (2) physical disintegration in response to internal or external stress, (3) mechanical abrasion and erosion by running water, and (4) biological attack by plants. The kind of weathering and erosion at any locale becomes a question of which of these processes dominate.

Of the common minerals of granite, hornblende weathers most rapidly, followed in order by biotite, plagioclase (albite), orthoclase (microcline), and quartz

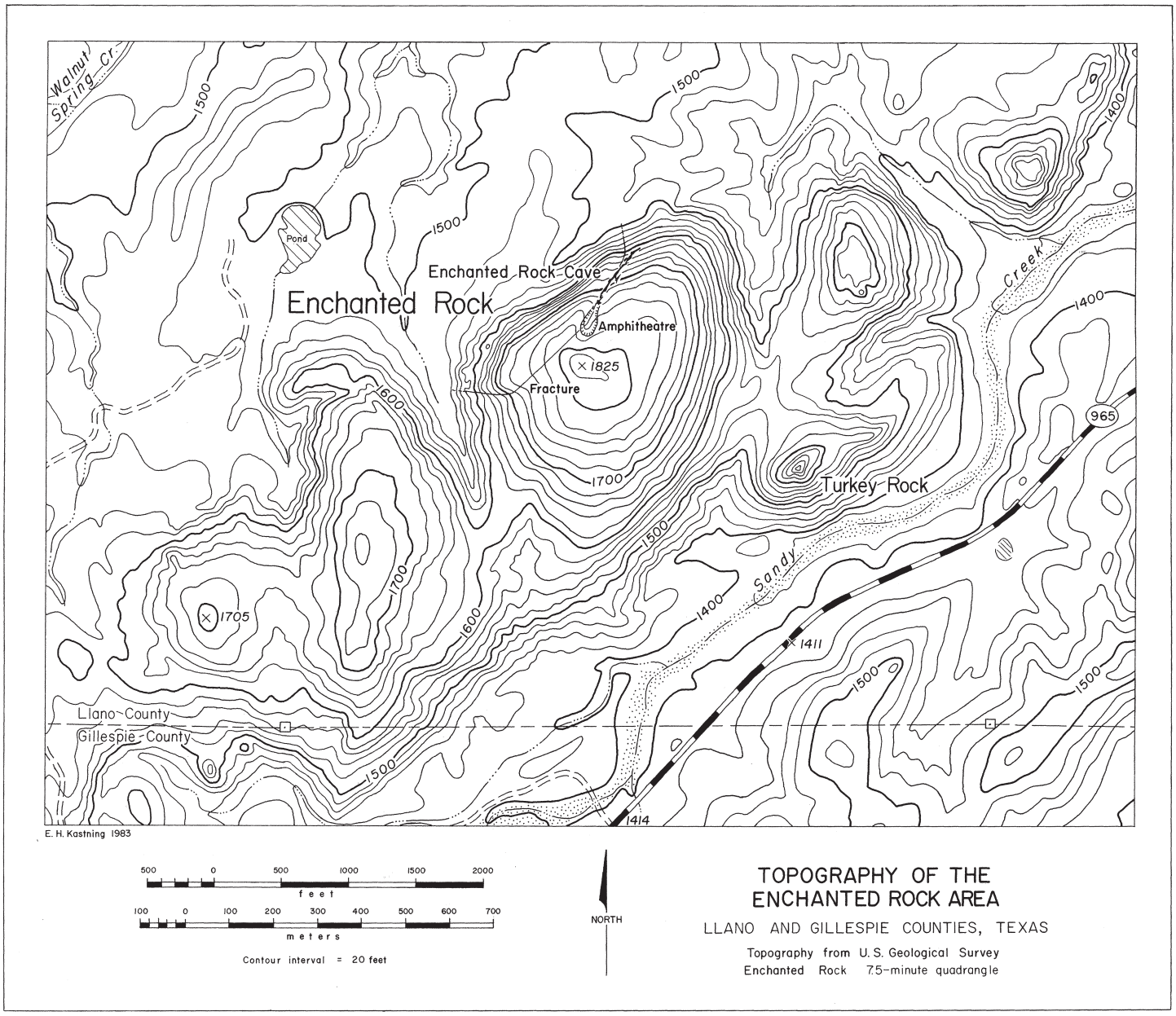


Figure 156: Topographic map of Enchanted Rock and vicinity, Llano and Gillespie Counties.

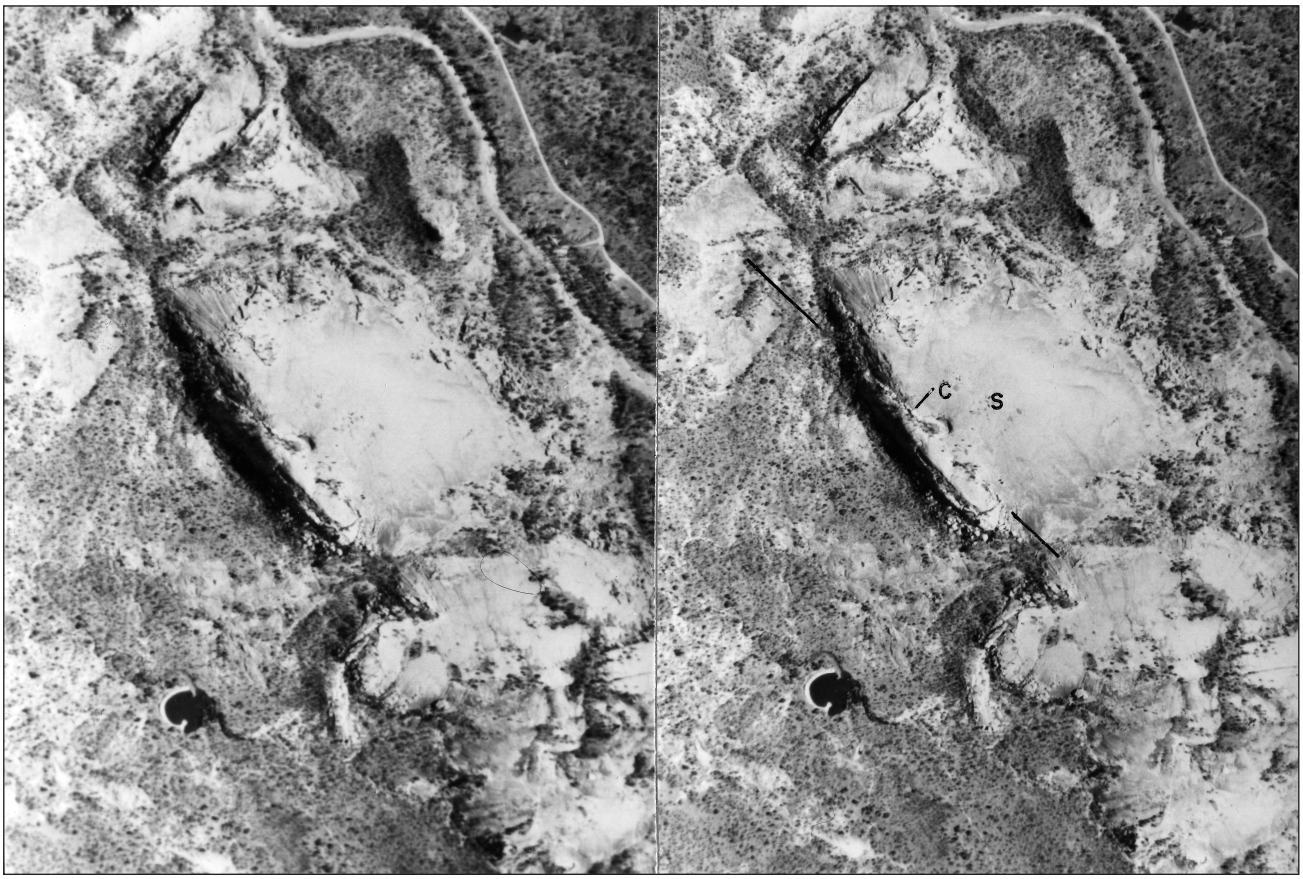


Figure 157: Stereoscopic aerial photographs of Enchanted Rock. The view is vertical and north is to the left. The summit is at "S" and the main entrance to Enchanted Rock Cave is at "C". The northeasternly trending fracture that is parallel to the steep, northwestern face of the dome and contains Enchanted Rock Cave, is indicated by the line. The sharp peak just above and to the right of Enchanted Rock is Turkey Rock. Stereoscopic viewing of these photographs shows significant vertical exaggeration. Photographs by U.S. Department of Agriculture (Photos DMH-8V-30, 31; taken 1 January 1958)

(Goldich, 1938). Biotite weathers rapidly owing to oxidation of the iron and expansion of the micaceous structure upon hydrolysis. Outward pressure from expansion of biotite tends to shatter adjoining grains of other minerals, contributing to granular disaggregation of granite into grus (Blackwelder, 1929; Dragovich; Clark, 1968; Eggler and others, 1969; Birkeland, 1974, p. 74, 138; Isherwood and Street, 1976). As chemical weathering progresses, hornblende, altered biotite, and plagioclase are removed, leaving behind granules of orthoclase and quartz (Melius, 1982).

Petrographic examination of samples of grus from the Llano Basin by E.B. Patton (1977), Begle (1978), Patton and Folk (1978) and Folk and Patton (1982) has shown that formation of grus occurs in two steps. First, weathering of granite is initiated through the development of small fractures parallel to the surface of weathering. These appear to be the result of upward arching of thin sheets (1 mm to 3 cm apart) either in response to release of compressive stress (Larsen, 1948; Twidale, 1962, 1964; Twidale and Sved, 1978), or as forces of expansion are confined laterally through buttressing (Begle, 1978; Patton and Folk, 1978; Folk and Patton, 1982). Expansion may be caused by erosional unloading, solar insolation, or incipient expansion of biotite. Secondly, because water can then seep into the shattered zone, the thin sheets of granite disintegrate along randomly oriented fractures, boundaries of grains, and planes of cleavage as a result of hydration and expansion of biotite. Therefore formation of grus is a physico-chemical process involving sheeting and chemical alteration.

Mechanical weathering by running water plays a secondary role in weathering of granites in central Texas. Grusified rock may be abraded and transported by sheetwash or by flow in small channels and runnels during runoff from storms (Hoskin and Sundeen, 1979). Removal of disaggregated grus allows the zone of weathering to penetrate further into bedrock.

Biologically induced weathering of granites has occurred at Enchanted Rock. A variety of plants have taken root in fractures and gnammas on the dome (Whitehouse, 1931). These include grasses, shrubs, cacti, and a small number of trees. Chelation and accumulation of silica by plants augments weathering and development of grus (Lovering, 1959). Lichens may locally contribute to weathering of granite as well (Fry, 1926, 1927; Hale, 1967; Scott, 1967; Syers and Iskandar, 1973).

Twidale and Bourne (1976) have shown that, in the presence of moisture, surfaces of exposed granite become pitted beneath a regolith of soil and grus. Chemical attack removes hornblende, biotite, and plagioclase, leaving behind small pits whose depths and widths are proportional to the grain-size of the bedrock. Coarse-grained, porphyritic, granitic rocks, such as those at Enchanted Rock, display pits ranging from 1 to 3 cm in depth and width. Pitting is also seen in channels on bedrock (gutters) and granitrillen, where loose grains have been plucked by running water

The sequence of weathering is also manifested in the textural and compositional variation of particles transported in local streams of the Llano Basin (Bradley, 1970; Hoskin and Sundeen, 1979). McEwen and others (1959) have shown that larger particles in upstream reaches in comparison to those in downstream reaches are more spherical and contain a greater proportion of multi-grain rock-fragments. As transportation increases, size of grains decreases and particles become less spherical owing to the angular nature of individual grains of ortho-clase. Small particles, such as grains of biotite and hornblende, are flushed downstream, and arkosic sands are left in beds of channels in the Llano Basin.

STRUCTURAL LANDFORMS

The largest granitic landforms of the Llano Basin are the large domes standing high in relief above the floor of the Llano Basin. Two varieties are found here. Rounded hills (bornhardts), such as Enchanted Rock, are massive and result primarily from exfoliation of outer sheets of rock along curvilinear joints which parallel the surfaces of the domes (Figures 157 and 159). Angular hills (castle koppies), such as Watch Mountain and Turkey Rock (Figure 157), are generally smaller and originate where weathering has separated blocks of granite along rectilinear joints. Castle-koppies are more susceptible to erosion because water can penetrate deeper into the rock along vertical joints. Blocks outlined by joints weather into corestones within a matrix of grus. Spherical boulders are left behind upon removal of the grus. Subsequent weathering leads to exfoliation of the outer, concentric shells of these boulders

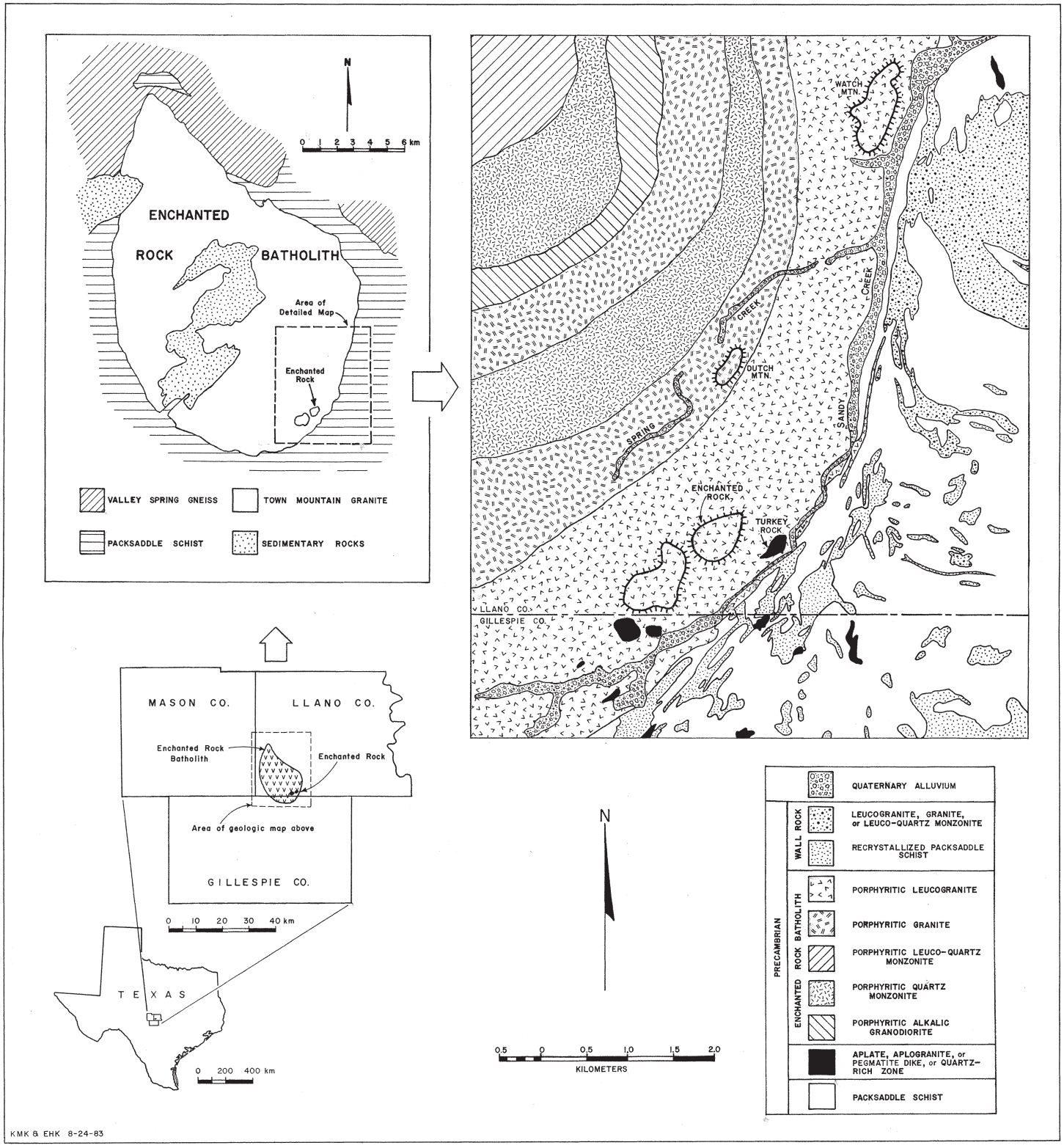


Figure 158: Geologic map of Enchanted Rock and vicinity. Modified from Hutchinson (1956, 1960) and Barnes (1952b).

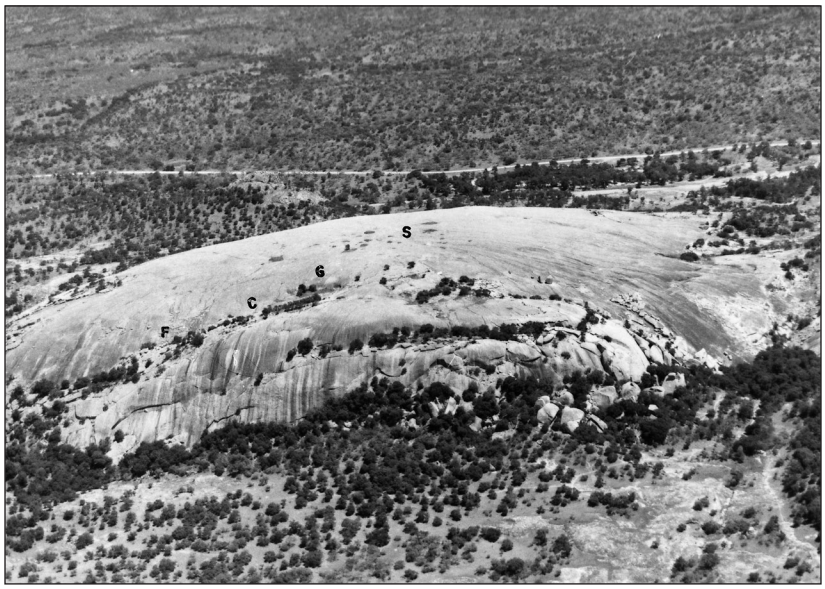


Figure 159: Oblique aerial photograph looking south at the northwestern side of Enchanted Rock. The summit is at "S". The large amphitheatre-like gnamma is at "G". Water is concentrated by this depression and directed into the large fracture "F" containing Enchanted Rock Cave. Entrance to the cave is at "C". Trees have taken root along traces of fractures.

(Chapman and Greenfield, 1949; Ollier, 1971).

There has been a great deal of controversy about whether bornhardts develop above or below the regolith (Bain, 1923; King, 1966, 1975, 1977; Twidale, 1964, 1981, 1982a,b; Twidale and Bourne, 1975a,b, 1977a; Fairbridge, 1977). The prevailing thought is that they are created at least partially in the subsurface. Rocks that weather deeply have a high density of joints and the joints penetrate to great depths. Some researchers have argued that highly jointed rocks become lowlying areas of an erosional plain, whereas relatively unjointed rocks eventually stand out above the plain as domes. However, the existence and location of domes may also be determined by variations in the original petrofabric developed in the igneous rock during cooling. On the other hand, domes may form in zones initially protected by an indurated, siliceous crust. Enchanted Rock and neighboring domes are located in the outer, porphyritic, leucogranitic zone of the batholith. This suggests control by the petrofabric. Some structural control in the immediate area of Enchanted Rock is evident from aerial photographs as well (Figure 157).

GRANITIC PSEUDOKARST

Many, minor landforms developed on Enchanted Rock are properly termed pseudokarstic as they resemble solutional forms. They are the result of physico-chemical processes other than dissolution, including grusification, granular disintegration, physical erosion, and the activity of plants. A few small caves owe most of their development to purely physical or mechanical processes.

Gnammas

Numerous shallow depressions ranging in diameter from 0.3 to 15 m occur on gentle slopes of Enchanted Rock, particularly near its summit (Figure 160). These features are called gnammas, but they have been variously described in the literature under many other names, including opferkessel, solution basins, depressions, pits or pans, rock pools, basins or tanks, charcos, tinajitas, kamenitzas, and etched potholes. Descriptions and reviews of depressions formed on granitic and other crystalline bedrock may be found in Smith (1941),Wentworth (1944), LeGrand (1952), Wilhelmy (1958), Hedges (1967, 1969), and Twidale (1982a). Morphologically gnammas may resemble true, karstic depressions formed on carbonate rock by dissolution, such as those commonly occurring in central and western Texas (Udden, 1925; Smith and Albritton, 1941).

Initial depressions on Enchanted Rock may have developed in any of several ways in accordance with processes described by Twidale and Corbin (1963). Concentrations of minerals most susceptible to weathering (e.g. biotite) may produce small depressions. Some depressions apparently have developed by differential weathering along joints (Figure 161).

Lichens may have been responsible for the onset of some gnammas. In other cases, the growth of roots entering the upper surfaces of the rock has dislodged particles of rock.

Apparently most gnammas at Enchanted Rock were initiated by flaking of the surface of the rock (Figure 162). As discussed previously, the uppermost zone of bedrock may become shattered by buttressed expansion owing to erosional unloading, solar insolation, or hydration of biotite. A thin sheet of granite may become detached as it arches upward. The "hollow" sound of parts of the granitic surface when walked on confirms the presence of these thin sheets. Chemical attack by aqueous seepage increases as the rock splits, and the arched sheet may then disintegrate into smaller fragments which, in turn, weather in place to finer debris (Figure 162).

The original, small depressions are enlarged by the corrosive activity of mildly acidic water which fills them after rains or by organic acids produced by lichens and algae. Feldspars are generally altered to kaolin, and biotite is changed to vermiculite (Twidale and Corbin, 1963). An abundance of large gnammas on gentle slopes suggests that water contained in these basins is the principal agent in their enlargement. Another indication is that bedrock-floors of gnammas are horizontal rather than parallel to sheeting joints

Gnammas on Enchanted Rock vary in size and in the amount of sediment they contain. Some are floored by bedrock, whereas others contain a blanket of disaggregated grus, sand, and in some, cobble-sized flakes of granite. Some gnammas have a blanket of detritus thick enough to harbor vegetation such as grasses, cacti, shrubs, and trees (Figures 160 and 161) (Whitehouse, 1931).

Gnammas on slight slopes are periodically flushed by heavy rains, but on level surfaces coarse material may remain as a residuum that is subject to further chemical attack. Finer material may be removed by deflation.

Rock Doughnut

Several gnamma-like features on Enchanted Rock are encircled by a raised rim of granite 20 to 30 cm in width and up to 15 cm high. These were originally described by Blank (1951b) who named them "rock doughnuts" and hypothesized that their origin may be attributable to any of three processes: (1) induration and case hardening of surrounding rock, (2) reduction of velocity and corrosive effect of sheetwash in the vicinity of water-filled gnammas, and (3) development of gnammas upon low, residual domes between gutters. The first two hypotheses are untenable, and Twidale and Bourne (1977b) have elaborated on the third and have presented the following alternative hypothesis: Raised platforms (plinths) may develop beneath boulders on a sloped surface as neighboring rock is weathered

Figure 160: Gnammas near the summit of Enchanted Rock that contain water from recent rain. Incised channels (gutters) such as the one in the foreground drain many gnammas and integrate runoff into networks of drainage. Grasses have taken root in gnamma at upper right.

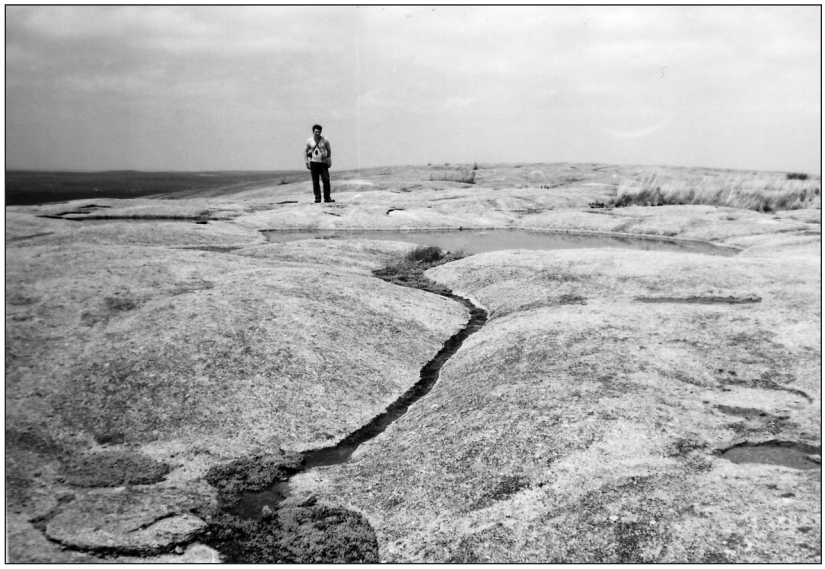




Figure 161: Gnammas developed along a joint on Enchanted Rock. Grasses and prickly-pear cactus have taken root in detritus.

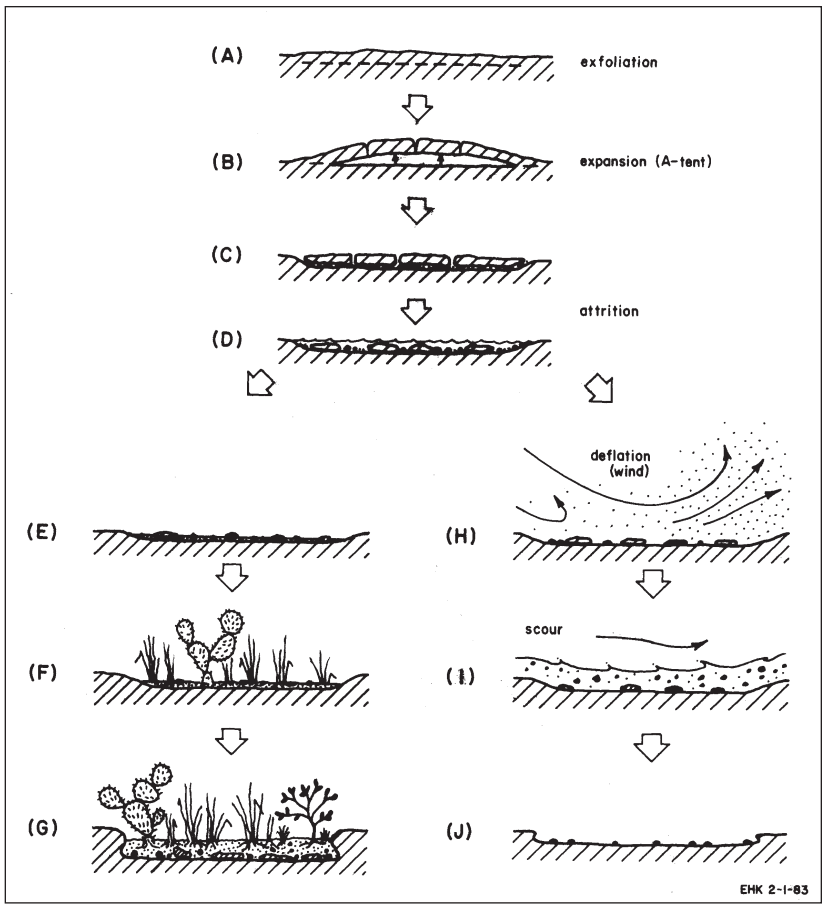


Figure 162: Origin of gnammas. (A) Surface of rock is fractured into thin sheets by release of internal, compressive stress, buttressed expansion, erosional unloading, solar insolation, or incipient expansion of biotite. (B) Sheet arches upward as a release of stress. (C) This sheet disintegrates along random fractures, grain boundaries, and planes of cleavage as a result of hydration of biotite and expansion. (D) Rainwater retained in gnamma attacks fragments which are reduced in size by attrition. (E) If protected from severe runoff or wind a residuum may develop. (F) Vegetation may take root in residuum. (G) Further physico-chemical and biological weathering will deepen gnammas and residuum. (H) Fine debris is lost by deflation where gnammas are exposed to wind. (I) Coarser debris may be flushed from gnammas during heavy runoff. (J) Rock-floored gnammas develop where water and wind have removed detritus. Modified from Twidale and Corbin (1963, p. 13) and Folk and Patton (1982).

Figure 163: Stains indicating seepage of water emerging from sheeting joints on Enchanted Rock.



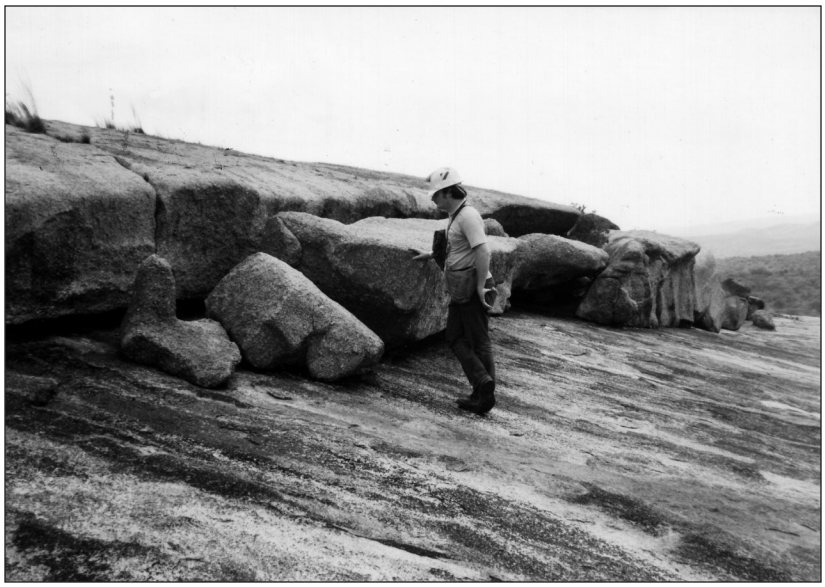


Figure 164: Tafoni (cavernous weathering) beneath exfoliated sheets on Enchanted Rock. As weathering progresses, blocks break away along fractures and are reduced in size.

away. Moisture retained between the plinth and boulder is protected from evaporation and begins to form a gnamma on the surface of the plinth. Upon eventual disintegration of the boulder the plinth and gnamma remain in the form of a rock doughnut. Rock doughnuts therefore develop as a result of local inversion of relief. They are ephemeral features, easily destroyed by surficial wash and scour (Twidale and Bourne, 1977b).

Granitrillen and Channel

Incised flutes and channels (granitrillen and gutters) are common on the surface of Enchanted Rock, particularly where drainage from gnammas has been integrated into networks of flow (Figure 160). These are similar to gutters and lapies found on inselbergs in Australia and elsewhere (Wilhelmy, 1958; Dragovich, 1968). Most channels on Enchanted Rock are several centimeters deep and 10 to 20 cm wide. Their beds are pitted, and quartz and microcline stand out in relief. Channels are only a few centimeters deep on gentle slopes, such as those near the summit of Enchanted Rock, but they are much more deeply incised, as deep as 100 cm, on steep slopes, apparently in response to steeper gradients. First-order channels converge downstream, forming higher-order channels that carry greater progressively amounts of runoff. As a result, widths of channels increase downstream. The relatively level area at the summit of Enchanted Rock has been etched into a fluviokarstic landscape by gnammas and networks of integrated channels. As will be seen below, one of these networks contributes significant runoff to Enchanted Rock Cave.

Tafoni, Arches, and Pedestals

Water running down the surface of a granitic dome during heavy rain may infiltrate fractures, especially those separating exfoliated sheets (LeGrand, 1949). This water then flows on the surface of the dome beneath these sheets and emerges at exposed margins of the sheets (Figure 163). The basal parts of the exfoliated blocks are physically and chemically weathered. Tafoni, or small caves, develop where blocks have been undercut at their contact with the bedrock (Figure 164) (Jennings, 1968; Dragovich, 1969; Grantz, 1976; Pestrong, 1979; Twidale, 1982a, p. 281, 283-300). Cavities result from differential weathering owing to microclimatological differences between the surface and underside of the blocks. Temperatures are more uniform and moisture is high beneath the blocks. Furthermore, induration and case-hardening protect the upper surfaces of the blocks, causing higher rates of weathering beneath the blocks. Hydrolysis, hydration, and weathering by salt of micaceous minerals lead to flaking and excavation (Dragovich, 1967; Evans, 1970; Martini, 1978; Bradley and others, 1978, 1980; Winkler, 1980).

Further weathering may result in arches and pedestals. The latter commonly consist of core-stones perched on bases of grus. These are similar to pedestal-rocks described by Bryan (1925, 1926), Leonard (1927), Petty (1932), Crickmay (1935), and Ollier (1978)

A-tents

In several places on Enchanted Rock, pairs of rectangular slabs have been forced upward into tent-shaped structures (Figure 165). These forms, known as A-tents, are similar to those on granitic outcrops in Texas, Australia, and elsewhere (Osmond, 1950; anonymous, 1951; Jennings and Twidale, 1971; Peterson, 1975; Jennings, 1978; Twidale and Sved, 1978; Twidale, 1982a, p. 317-329). Jennings and Twidale (1971) and Twidale and Sved (1978) suggest that exfoliated slabs are pushed upward in response to release of compressive forces within the mass of the dome. Folk and Patton (1982), on the other hand, suggest that slabs buckle by buttressed expansion in response to erosional unloading, solar insolation, or chemical alteration of biotite. On steeper slopes on Enchanted Rock it is possible that arrest of downslope movement of exfoliated sheets may have caused, in itself, sufficient compressional force to produce some A-tents. The small caves formed in this manner are pseudokarstic and of purely physical origin.

Caves Formed by Exfoliation and in Talus

Numerous caves exist around the peripheries of Enchanted Rock and neighboring domes that are merely spaces beneath dislodged plates, blocks, and boulders. Sufficient separation and sliding of exfoliated sheets from the surface of the dome may result in small cavities. In such cases the sheets come to rest over channels or hollows on the dome or lean across lines of inflection where steep surfaces on the dome level off. Caves of this variety are common in many areas where slabs have moved to the bases of domes or have bridged and roofed crevices, depressions, or talus (Redden, 1963; Twidale, 1982a, p. 322-329).

Figure 165: Stereoscopic view of exfoliated sheets on Enchanted Rock that have been pushed upward by lateral compression into A-tents. The small cave formed beneath this A-tent is 0.6 m high. This structure has formed in place, without downslope movement of the sheets.



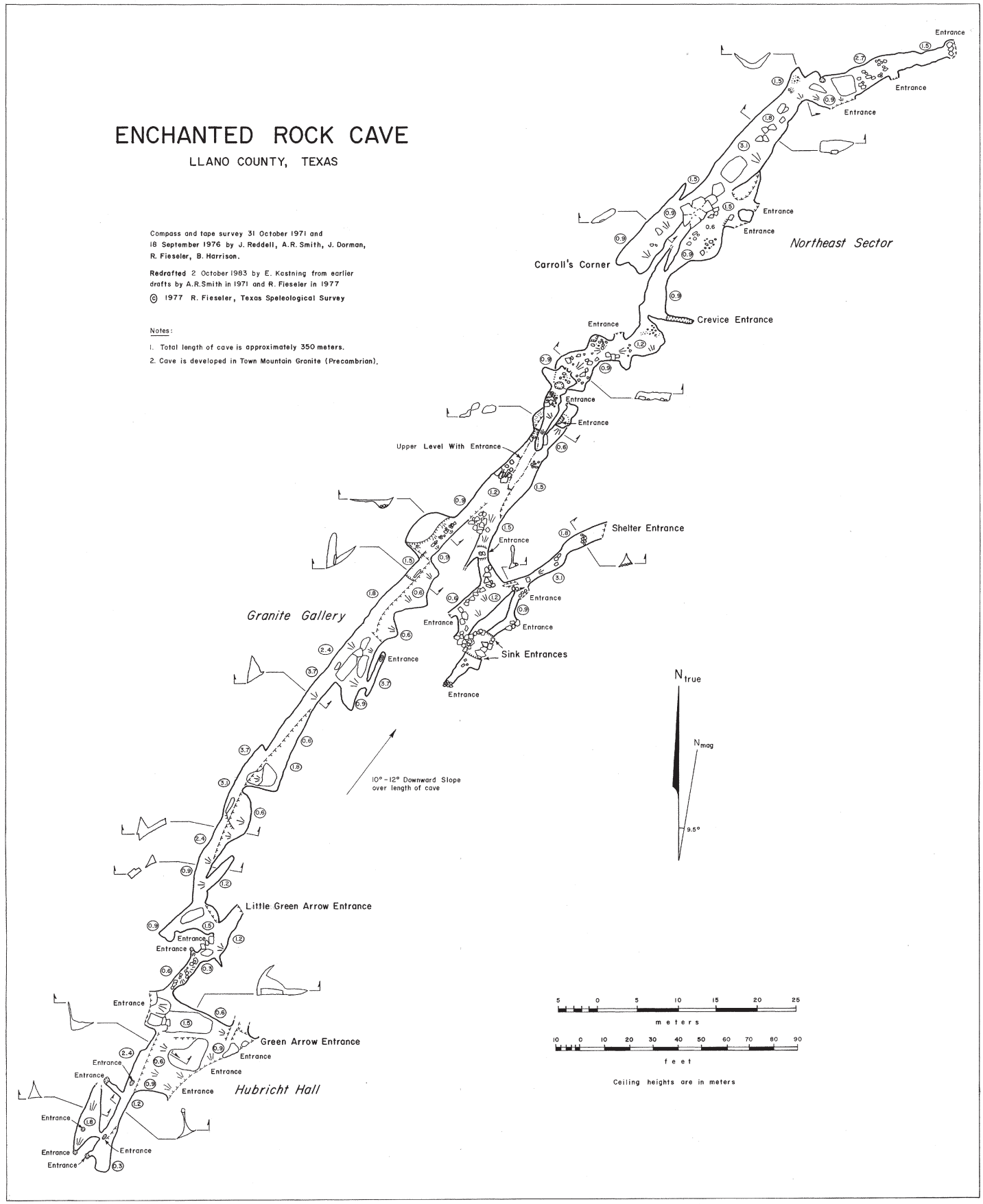


Figure 166: Map of Enchanted Rock Cave. Modified from Smith (1974) and Fieseler and others (1978, p. 34).

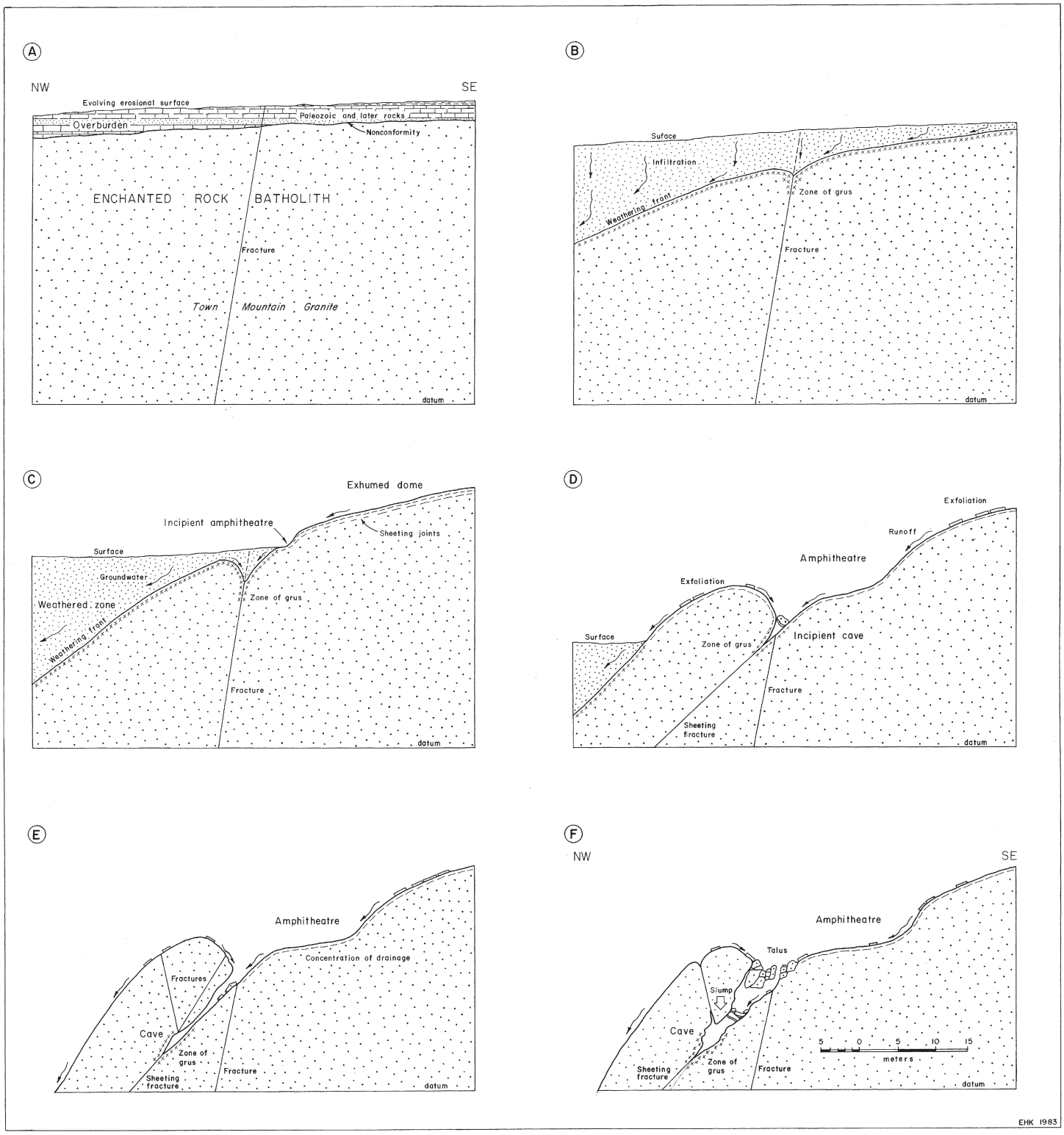


Figure 167: Diagrammatic summary of the inferred evolution of Enchanted Rock Cave. (A) Batholith is buried; fracturing occurred during cooling. (B) Overburden is removed by erosion and chemically aggressive, seeping water begins to attack the granitic surface and fractures. (C) Subsurficial erosion continues and rounding of dome becomes pronounced where highly fractured areas are more easily weathered to grus. The dome becomes exposed to subaerial weathering. (D) Runoff is channeled into gutters and along the host-fracture. The large amphitheatre-like gnamma develops and concentrates runoff into the fracture. (E) Fracture is enlarged by grusification and mechanical suffosion. Note zone of grus above where fracture is still closed. (F) Cave becomes roofed by displaced blocks and exfoliated debris shed from surface of dome. This is the present stage of development of the cave.

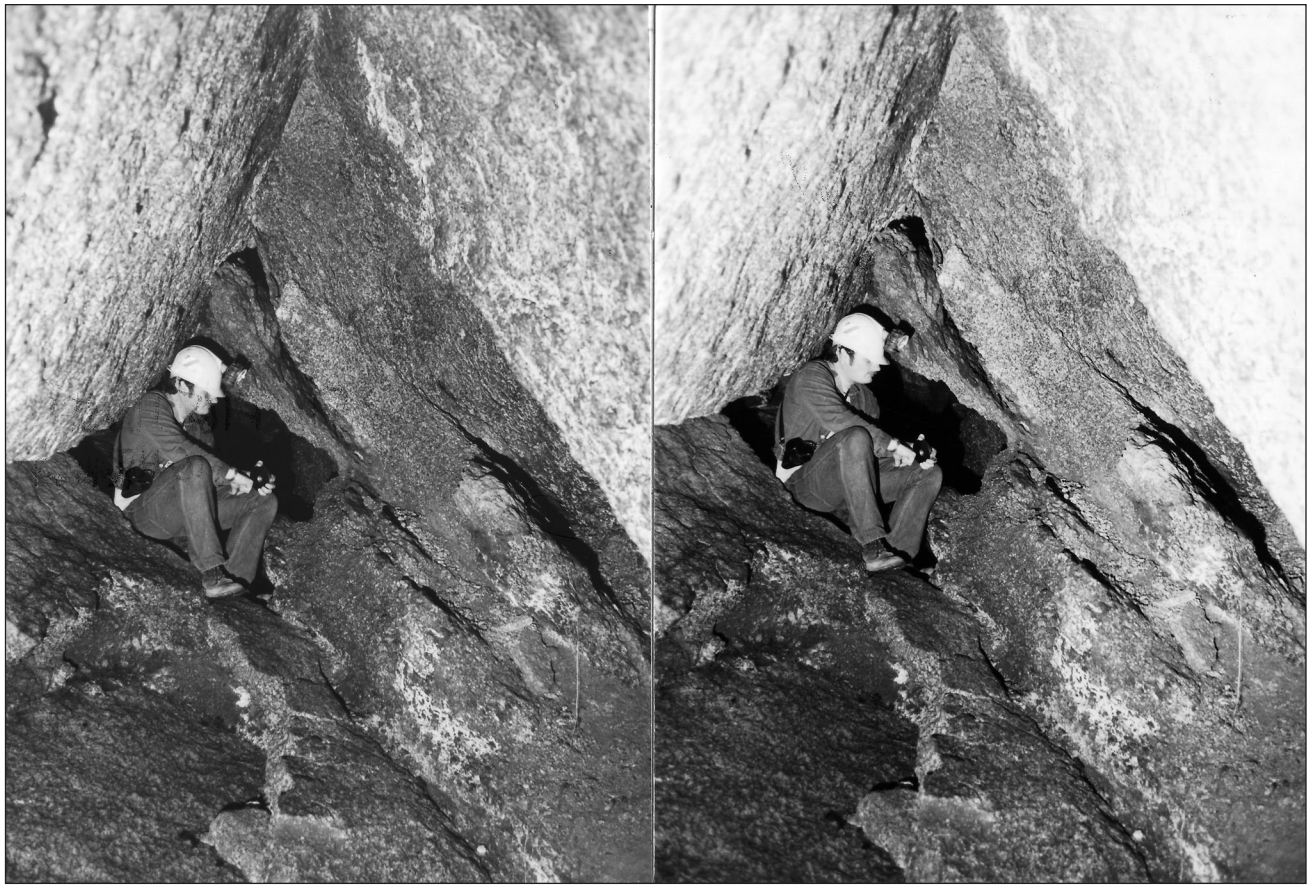


Figure 168: Stereoscopic view of interior of Enchanted Rock Cave at the southwestern end of the lower level of the main passage (Granite Gallery). View is to the southwest. A 2 m-thick zone of grus is visible on the right wall and floor. The passage has formed where grus has been removed along the large, sheeting joint (at the floor) and a near-vertical fracture (at top of photograph).

Blocks, boulders, slabs, and plates may easily crack and split along fractures (Twidale, 1982a, p. 301-307). As this happens, slumping and other displacement may allow the pieces to fall together or lean against neighboring boulders and rocks and thereby form small caves. Such openings are common within accumulations of talus at the bases of Enchanted Rock and other domes in the vicinity.

ENCHANTED ROCK CAVE

Enchanted Rock Cave (Figure 166) is one of the largest caves known to have developed within the mass of a granitic dome. Its origin, as described by Kastning (1977a,b, 1979), is attributed to four processes: (1) fracturing of the bedrock, (2) grusification, (3) removal of detritus by mechanical suffosion, and (4) roofing by accumulation of talus. Boone's Cave, a smaller (40 m long) cave in granite-gneiss in North Carolina (Hedges, 1978), and several in California are similar

Fracturing

Two distinct sets of joints occur in the vicinity of Enchanted Rock (Hutchinson, 1956, plate 1). One set trends N 40°-60° E and the other, orthogonal to the first trends N 30°-50° W. Hutchinson (1956) showed that these joints result from stresses

Figure 169: Panoramic view of large amphitheatre-like gnamma below summit of Enchanted Rock. Runoff concentrated by this feature drains to the right and into fracture at trees in center-right of photograph. The entrance of Enchanted Rock is among boulders at right.



incurred during emplacement of the batholith.

A prominent, near-vertical joint, extending approximately 500 m N 47° E, runs through Enchanted Rock and parallel to its northwestern face (Figures 157 and 159). Because this fracture corresponds with the N 40°-60° E set of joints of Hutchinson (1956), it presumably formed during granitic intrusion and thereby predates exhumation of the dome (Figure 167a).

A sheeting joint, dipping approximately 45° to the northwest, separates the large plate forming the northwestern face of Enchanted Rock from the mass of the dome. Enchanted Rock Cave has developed along the intersection of this fracture with the near-vertical joint.

Grusification

Enchanted Rock Cave has been enlarged through grusification. Chemical alteration of the walls of fractures within Enchanted Rock may have begun prior to exhumation of the incipient dome (Figure 167B). Aqueous seepage probably entered along both the near-vertical joint and the sheeting joint. Water coming in contact with the walls of these joints accelerated the weathering process, particularly along the intersection of the two. This has resulted in a zone of grus that is now up to 1 m thick in this vicinity (Figure 168). Below this zone fractures remained tightly closed and there has been no grusification. Seepage that reached this depth flowed laterally to the northeast along the line of intersection of the two joints. Following rainstorms some water undoubtedly remained perched in pools at this level, allowing grusification to proceed more rapidly along the walls and floor (Figure 167C). Today, pools can be found in the cave following heavy rains.

A large gnamma has formed on the surface of the dome, adjacent to and southeast of the near-vertical fracture (Figures 167D). This feature began perhaps as a small gnamma and has subsequently enlarged upslope and away from the fracture by grusification and exfoliation along its periphery. At approximately 20 m in diameter it is presently the largest depression on Enchanted Rock (Figures 159 and 169).

Morphologically and genetically it is similar to the armchair- and canoe-types of gnamma of Twidale and Corbin (1963). The effect of this feature on the origin of Enchanted Rock Cave has been profound. Runoff from a large area near the summit of Enchanted Rock has been concentrated by this basin and channeled directly into the main fractures of the cave. During severe storms runoff has undoubtedly been very high and has probably accelerated grusification and erosion substantially.

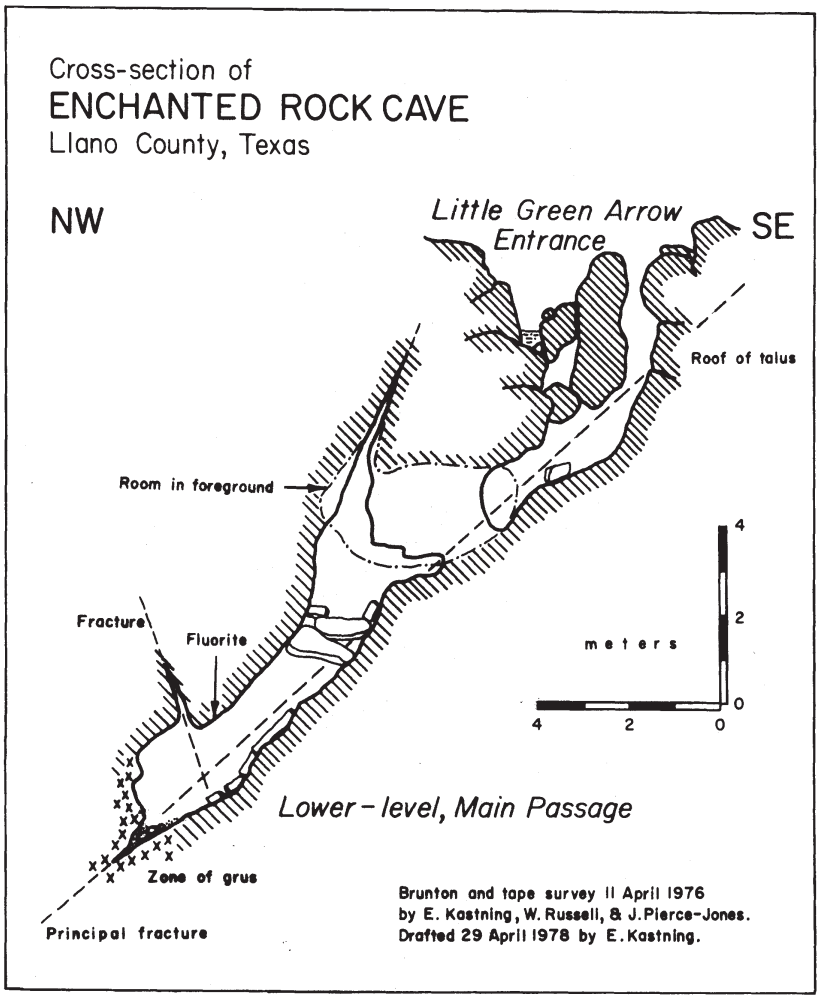


Figure 170: Surveyed cross-section through Enchanted Rock Cave at the Little Green Arrow Entrance (see Figure 166). The passage on the lower level (Granite Gallery) has developed in grus and the upper levels are roofed by talus.

Mechanical Suffosion

Once hornblende, biotite, and plagioclase were removed from the zones of grus within fractures, the remaining grains of quartz and orthoclase easily disaggregated and spalled from the walls, forming a deposit of gravel on the floor of the cave. Some of this material has been flushed from the cave from time to time by catastrophic runoff. The removal of detritus from the subsurface, referred to as mechanical suffosion or piping, is still operating in Enchanted Rock Cave.

In places widths of fractures have enlarged to over 2 m by grusification, spalling, and mechanical suffosion, forming the main passages in the cave (Figures 166, 167E, and 168). Once continuous passage, 2 to 3 m high and 1 to 3 m wide, follows the intersection of the near-vertical and sheeting joints for over 80 m. Deposits of gravel cover much of the floor, but in places it is possible to see a continuous span of bedrock from the floor, along the walls, and to the ceiling of the cave (Figure 168).

Roofing by Accumulation of Talus

The enlarged fractures were probably open to the surface at one time. Large granitic blocks, that have been weathered out along smaller joints and fractures, have since been dislodged from the upper walls of the fracture and displaced downward into the fracture (Figure 167F). Moreover, some slabs of exfoliated sheets from the locally steep surface of the dome have accumulated within the fracture, sealing many of the interstices among larger blocks. Much of the cave became roofed in this manner. Figure 170 is a cross-section of the cave as it appears today, showing the passage on the lower level developed in grus and upper levels roofed by talus.

Enchanted Rock Cave has approximately 350 m of surveyed passage (Figure 166). Most of the cave is within a debris of talus, and there are at least 30 entrances. Maximum relief in the cave is approximately 30 m. The highest point is at the southwestern end, and the cave generally slopes downward toward the northeast. However, the main passage in the zone of grus is nearly horizontal. An intermittent spring, issuing from the talus near the base of Enchanted Rock northeast of the cave, is believed to be the point of discharge for waters flowing through the cave during wet weather.

SUMMARY

Weathering has produced several interesting, pseudokarstic phenomena on granite in central Texas. Gnammas, rock doughnuts, and granitrillen have been etched into the smooth granitic surfaces, forming a miniature fluviokarst. Small caves in the form of tafoni, arches, and A-tents, which are relatively rare in the United States, are found there. Enchanted Rock Cave has evolved from a fortuitous, geologic setting and a sequence of events that includes fracturing, concentration of drainage by a large gnamma, grusification, mechanical suffosion, and roofing by accumulation of talus.

Many small granitic landforms discussed in this chapter are younger than the period of denudation which removed overlying rocks from the granite. Consequently, they have evolved since the Miocene Epoch. Twidale and Bourne (1975b) have shown that minor granitic landforms may be initiated beneath a regolith, in which case some of the gnammas, channels, and tafoni on Enchanted Rock may have been initiated in the Tertiary. Many surficial features may be Pleistocene or Holocene. However, Enchanted Rock Cave, which began as the dome was being exhumed, may in part be older, perhaps initiated during the late Tertiary. Some of the processes which formed it are still active today.

PART III

SYNTHESIS, CONCLUSIONS, AND RECOMMENDATIONS

14
REGIONAL SYSTHESIS AND CONCLUSIONS

INTRODUCTION

Detailed investigations of nine individual areas of karst development within the Edwards Plateau (Chapters 5 through 13) show that evolved landforms are diverse in morphology and that controls active during their geomorphic evolution have varied considerably over the region. Not only have the types of controlling factors differed among the chosen areas, but so have the intensity and duration with which they have operated. Moreover, the specific factors which guided karstic processes have interacted among themselves. In some cases they acted contemporaneously, resulting in a superposition of morphogenetic characteristics in the produced landforms. In other cases processes and controlling conditions followed in chronologic sequence, giving rise to progressive generations of landforms. In situations where controls were numerous, equally dominant, or complexly interrelated, geomorphic analysis has proven difficult and has necessitated detailed examination of all relevant factors.

What follows is a review of the findings of the areal studies as developed in the preceding chapters. Thus, the geomorphic histories of karst development within individual areas are integrated into a regional model.

CONTROLS ON DEVELOPMENT OF CAVES

Lithostratigraphic Factors

Aside from some of the Precambrian and Paleozoic rocks of the Llano Basin, nearly all lithic units exposed on the Edwards Plateau and those forming the major aquifers of this region consist of carbonate materials. Most are limestone, dolostone, or marl. Carbonate rocks are typically much more soluble in slightly acidic meteoric waters than are most other types of sedimentary rocks (with the exception of certain evaporitic materials, such as gypsum and rock salt). Moreover, highly soluble rocks, such as very pure calcareous beds, are karsted and cavernous in comparison to adjacent, superjacent, subjacent, or nearby units of relatively low resistance to dissolution owing to low content of calcite (e.g. dolostone, marl, argillaceous or arenaceous limestone). Consequently, development of karst and caves in regions of diverse types of rock is highly localized and confined to particular exposures on the surface, facies, or beds within stratigraphic sequences.

The lithic character of the rock within each investigated area on the Edwards Plateau has been shown to be an important factor in guiding dissolutional enlargement. For the most part passages in caves, or segments thereof, are conformable to stratification. Although in some cases this reflects circulation and dissolution along horizontal zones of relatively high permeability (e.g. bedding-plane partings), there are sufficient examples among caves discussed in Part II to support the conclusion that conduits and chambers formed within lithic units that are most susceptible to dissolution. In general, such rocks (1) are high (greater than 90 percent) in calcite content, (2) have relatively well developed primary porosity (e.g. burrows), (3) have undergone diagenetic changes (e.g. leaching, dedolomitization, cracking) leading to enhancement of secondary porosity, or (4) are characterized by two or more of these attributes.

With the exception of deep caves in the Langtry area, passages in caves of the Edwards Plateau are generally horizontal or subhorizontal and concordant with lithologically favorable strata. In the Langtry area, however, the Devils River Formation is massive and vertically homogeneous, and passages have developed equally well at various depths.

Strata which are resistant to dissolution or erosion, in comparison to rocks above or below, commonly serve as caprocks over cavernous beds. In many cases capping beds have protected underlying caves from premature collapse and truncation. This has been particularly important in the development and preservation of caves in the Sonora and New Braunfels areas.

Structural Factors

Geologic structure can perhaps be considered the most fundamental of all factors which contributed to the development and modification of caves in the Edwards Plateau. This is as expected because the structural setting determines the degree of interaction among the various components of the evolving karstic landscape, including (1) hydrochemical processes that excavate openings in the rock, (2) fractures and other zones of secondary porosity that serve as avenues for circulation of groundwater, (3) integrated networks of flow that convey subsurficial drainage, and (4) lithostratigraphic zones of relatively high solubility. Furthermore, structural conditions dictate the positional relationships of significant lithic units with respect to prevailing flowpaths of groundwater and to the topographic setting.

The dominance of geologic structure in guiding cavern development became clear during the early phases of this research, and structure has proven to be the single factor that best explains morphologic differences among caves from the eight selected areas of investigation. Deformational features of the areas vary and reflect the underlying structural elements and regional tectonism that preceded cavern development.

The relationship of speleogenesis to structural setting has been discussed in considerable detail and summarized for each area (Chapters 5 through 12). However, some general conclusions can be drawn concerning the influence of structure on the origin of caves in the Edwards Plateau.

Attitude of Beds

The beds of carbonate units within nearly every cavern-bearing locality have a non-horizontal dip. Although in most cases the angle of dip is very low (commonly less than 2°) it is rarely equal to zero. But, even very low inclinations of beds influence the direction of migration of groundwater.

In general, excavation of horizontal conduits within the shallow-phreatic zone occurs parallel to the stratigraphic strike, along and just beneath the line of intersection between the water table and a particular, soluble and permeable stratum. Most caves of the Edwards Plateau which consist of tubular, master conduits of large cross-sectional area have developed along strike. This is indicated by horizontal gradients of such passages over long distances and by conformity of conduits to stratigraphic horizons.

In contrast, passages oriented along the direction of dip generally formed under vadose conditions, or in isolated cases where flow was guided deeply into the phreatic zone by confining beds or through relatively few open fractures (bathyphreatic conditions). Most dip-oriented passages are canyon-like in cross-section and once served as tributaries to strike-oriented, master conduits. Intersections of dip-oriented vadose feeders with strike-oriented phreatic conduits generally mark the position of the water table during the time of excavation of passages.

Fractures

Most caves investigated in this study were strongly influenced by fractures present prior to the onset of dissolutional excavation. Faults, joints, and bedding-plane partings constitute the initial openings through which water migrated, whether in the phreatic or vadose zone.

Observations of fractures within caves and analysis of maps of accurately surveyed caves established that linear segments of passages largely follow joints and more rarely faults. Most fractures are discordant with the strata and have vertical or near vertical dips. Linear segments of passages which are high and narrow tend to follow vertical fractures for greater distances than do low and wide segments. In general, the former tend to be conduits in the vadose zone which have developed down the dip, whereas the latter are typically avenues within the phreatic zone and along the strike of the strata. Segments of vadose conduits are

generally straighter in plan-view than those of phreatic conduits. This suggests that within the nearly horizontal strata of the Edwards Plateau, phreatic passages meander slightly along bedding-plane partings but are guided in part by vertical fractures. In many situations, passages that evolved in the shallow-phreatic zone are positioned along lines of intersection between permeable bedding-plane partings and vertical fractures.

It has been graphically demonstrated in the rose diagrams of the foregoing chapters that caves, and therefore fractures, lie along preferred orientations. Comparisons of frequencies of such orientations with trends of regional structural elements show that fractures are highly correlative with the tectonic history of central Texas as documented in the geologic literature. In some cases systematic sets of fractures created by temporally distinct episodes of deformation have been superimposed. Under such conditions, configurations of caves are notably more complex.

Spacing of fractures within an area has been a significant factor in the evolution of conduits. Where fractures were few and spaced far apart, flowpaths tended to be singular and had few tributaries. Moreover, in places of accentuated dip (e.g. the Devils River Uplift, Langtry area) flow was guided to great depths (i.e. bathyphreatic development). Conversely, in areas of higher densities of fractures, passages tended to be considerably more dendritic, with numerous small channels contributing flow to large, master conduits (e.g. Inner Space Cavern, Georgetown area). Even where dips were greater than average (e.g. greater than 2°) shallow-phreatic flow was still maintained.

A significant finding of this study is the degree to which the openness of fractures influences dissolutional enlargement of conduits. By assuming that tensionally produced fractures are generally more open than those created under compression, it has been possible to relate orientations of conduits to directions of previous tectonic stress. As a rule in areas of uniform geologic conditions, excavation of conduits is enhanced along open fractures produced by extension. This hypothesis is supported by the statistical proliferation of large passages along open fractures that correspond to known extensional stresses. Fewer passages formed along closed fractures of compressional origin. Further support is given by the character of fractures visible at the surface. For example, alignments of sparse vegetation taking root in joints correspond to tensionally produced fractures, whereas little vegetation is present along closed fractures. These findings are particularly well illustrated within the Langtry, Bend, Burnet, and Georgetown areas (Chapters 5, 8, 9, and 12 respectively).

Localization of Development of Caves

Explored caves tend to be clustered in specific areas within the Edwards Plateau. This observation, of course, led to the original selection of areas for detailed investigation. Comparison of distribution of caves, particularly the longest and deepest ones, with the structural elements of central Texas shows that caves were formed more readily in areas where tectonism has augmented the circulation of karstic groundwater. Three tectonic modifications to the geologic setting are dominant over the region. First, stresses incurred during uplift, subsidence, and broad regional warping produced extensive changes in the attitude of bedding. Second, these same forces enhanced the porosity of carbonate aquifers through intense fracturing. Third, differential uplift -- most importantly, elevation of the Edwards Plateau above the level of the land to the southeast -- has significantly heightened its gradients and avenues for migration of water for the establishment of conduit type of flow.

Structural Stability and Collapse

Fracturing weakens carbonate rocks as well as renders them more permeable. Gradual destruction of caves by collapse of overlying strata into cavernous chambers and concomitant stoping is common throughout the Edwards Plateau. Thin, brittle beds collapse more readily than do thick units as a result of closely spaced bedding-plane partings. However, the entire process is augmented where the rock has been weakened by closely spaced, vertical fractures. Additionally, intersections of major fractures characteristically promote the development of sizable rooms floored by large cones of collapsed debris. In many cases, collapse blocked major conduits that had actively transmitted groundwater and thereby led to flooding and development of large maze-like sections and routes of diversion around the debris.

Hydrogeologic Factors

Caves of the Edwards Plateau exemplify speleogenesis under the four most common hydrogeologic conditions: (1) flow just beneath the piezometric surface in the uppermost part of the phreatic zone (shallow-phreatic development), (2) circulation deep within the fractured phreatic zone (bathyphreatic development), (3) flow of streams and seepage through openings in the unsaturated zone above the piezometric surface (vadose development), and (4) periodic inundation and enlargement of previously formed passages within the lower vadose zone (development by floodwater).

Shallow-phreatic development is characterized by lengthy master conduits that are tubular, nearly horizontal, and concordant with stratification. The main passages of Caverns of Sonora (Chapter 6), Powell-Neel Cave System (Chapter 7), Gorman Cave (Chapter 8), Longhorn Cavern (Chapter 9), Natural Bridge Caverns (Chapter 10), and Inner Space Cavern (Chapter 12) are excellent examples. Most caves such as these have matured sufficiently to form networks consisting of the master conduit and several tributaries.

Bathyphreatic development generally forms deep caves that include vertical or steeply sloping passages. Water is typically force to depth along few vertically extensive fractures within a relatively homogeneous, stratigraphic sequence. Topographic relief of the Langtry area (Chapter 5) was sufficient to promote deep circulation along these avenues.

Vadose development of passages is largely a secondary and later modification to caves formed under phreatic conditions. Vadose conduits generally serve as tributaries to the phreatic passages and are graded toward them from points of recharge on the surface. At least some caves in each of the eight investigated areas (Chapters 5 through 12). Exhibit passages of this origin.

Modification of passages by floodwater is pronounced throughout the Edwards Plateau. Two lines of evidence show that many caves of the region were reflooded in the later stages of their development. First, maze-like passages are common in caves (e.g. Caverns of Sonora, Powell-Neel Cave System, Airman Cave, Inner Space Cavern). In some cases mazes formed where the main routes of drainage were blocked by collapse. In others, mazes were excavated by backflooding of caves during high stages of flow in nearby surficial streams. Second, redissolved speleothems in caves throughout the region show that reflooding was common, even among relict caves that are now considerably above local baselevels. This strongly suggests that climates wetter and cooler than those at present once existed in central Texas.

Topographic Factors

Development of caves is intimately associated with the geomorphic history of the landscape under which they form. The areal topography influences where water enters carbonate rocks and where it discharges. Moreover, hydraulic gradients are largely determined by differences in topographic relief in the vicinity of the evolving flowpaths within the aquifers.

Because much of the Edwards Plateau has been highly dissected by surficial drainage, local relief is typically high. In the Langtry area (Chapter 5) this has promoted deep hydraulic gradients and bathyphreatic development of caves. In many areas incision of surficial streams was so rapid that flowpaths and conduits within the aquifer did not have an opportunity to mature into well integrated systems. Consequently where the piezometric surface dropped rapidly in response to the cutting of valleys, caves were soon drained of water and left behind as isolated, dry voids. Such relict caves (known as “jug caves”) are merely chambers of various dimensions that have little or no interconnectivity. This variety of cave is the most common throughout the area of study.

Mature systems of caves -- those in which passages are long, of large cross-sectional area, near base-level, and well integrates -- have formed in places where dissection and valley downcutting was arrested for some time or where dissection has progressed to sufficiently low elevations whereby groundwater became perched on rocks resistant to dissolution. These conditions were prevalent along the Balcones Escarpment at the southern and eastern margins of the plateau and adjacent to major streams that traverse the plateau. Many passages of caves of the Langtry, Sonora, Burnet, Austin-San Marcos, and Georgetown areas, for example, are correlative with baselevels maintained over sufficiently long time to allow large conduits on distinct levels to form under shallow-phreatic conditions. Elsewhere, such as in the Menard, Bend, and New Braunfels areas, lengthy caves within highly soluble beds are perched above dolomitic, marly, or arenaceous strata. Many caves consist of multiple levels (e.g. Caverns of Sonora, Felton Cave, Longhorn Cavern, and Natural Bridge Caverns). In some of these, levels correspond at least in part to relatively static positions of the piezometric surface over intervals of time sufficiently long enough to allow maturation of master conduits and their tributaries.

Climatic Factors

Climate is an important aspect in the geomorphic evolution of karst because it directly governs the availability and quality of water flowing over, into, and through

the soluble rocks. Generally speaking, climate may not always determine which karstic landforms will prevail in a particular region (e.g. caves, dolines) or which processes will or will not occur. However, climate does govern the rates at which the processes operate. Climatic change in central Texas has modified many caves. Redissolved speleothems in relict caves now well above local baselevels indicate that precipitation, surficial runoff, and flow of groundwater increased sometime during the interval between initial enlargement of the case and the onset of the present-day climate. Reflooding of caves during wet climates is the most plausible explanation of region-wide redissolution of speleothems in caves no longer having sufficient inflow.

Many large sections of mazes in caves originated from flooding induced by collapse and blockage of active watercourses. However, it is likely that such enlargement by floodwater may have been accelerated by increases in discharge owing to wetter climates.

It is suggested that the wetter and cooler climates responsible for incursion of floodwaters into otherwise dry caves in the uplands of the plateau were related to periods of glacial advances in the American midwest. This hypothesis remains to be tested and correlated with the Pleistocene history of adjacent geographic regions such as the southern High Plains and the Coastal Plains of Texas.

CHRONOLOGY OF DEVELOPMENT OF CAVES

It is difficult to determine absolute ages for the development of particular caves or for stages of evolution of karstic aquifers in the Edwards Plateau. However, it is possible to establish limits on the intervals of speleogenesis, to state relative ages of episodes, and to place events into chronologic sequences. The discussion of each chosen area of investigation (Chapters 5 through 13) ends with such a synthesis.

Throughout the plateau there are numerous examples of karst that predate the most recent phase of speleogenesis. Paleokarstic phenomena include (1) collapsed and filled dolines from the Paleozoic in the Llano Basin (e.g. those of Chapter 8). (2) large-scale zones of collapse and subsidence in the Cretaceous rocks in the southwestern part of the plateau (Chapter 5), (3) vugular solutional cavities developed during the middle of the Cretaceous in response to uplift along the San Marcos Arch (Chapter 10), and (4) solutional-collapsed breccia associated with removal of gypsum in the Edwards Group of the central part of the Edwards Plateau.

The present phase of speleogenesis over the Edwards Plateau commenced with dissection of its surface by incision of surficial streams. This followed regional uplift of the plateau and creation of the Balcones Escarpment during the Miocene.

Evidence presented in Chapters 5 through 13 supports the premise that even though speleogenesis may have been initiated during the late Tertiary at some localities, caves were largely excavated during the Quaternary as the major streams cut deeply into the uplands. In fact, considerable enlargement may be attributed to the late Pleistocene when wetter and colder climates promoted flooding of many caves over the entire region. Large, well integrated conduits containing active streams, presently positioned near baselevel, and at grade to incised surficial drainage suggest that these particular caves are of relatively recent origin.

15
RECOMMENDATIONS FOR FURTHER STUDY

> Men occasionally stumble over the truth, but most of them pick themselves up and hurry off as if nothing has happened.
>
> —Winston Churchill
>
> Even if you're on the right track, you'll get run over if you fall asleep there.
>
> —Will Rogers

This study has investigated numerous karstic phenomena, some of which were previously unknown or undescribed, and has developed plausible hypotheses to explain the processes relevant to their origin. The research has attempted, with moderate success, to answer questions raised in prior investigations. As is the case with most efforts of scientific inquiry, the resolution of a particular problem or set of problems precipitates new questions which then beckon future attention. This chapter enumerates some of the unresolved matters and new issues stemming from this study.

REGIONAL SYNTHESIS OF LOCALLY DISPARATE CONDITIONS

> When we try to pick out anything by itself, we find it hitched to everything else in the universe.
>
> —John Muir

This constitutes the first systematic study of the hydrogeomorphology of the Edwards Plateau karst. Emphasis was placed on the distribution, morphology, and origin of cavities and conduits developed through dissolution. As summarized in Chapter 14, genetic processes responsible for development of caves have been controlled to various degrees by the topographic, geologic, hydrologic, and climatic settings of the region. In some cases, controls have operated on a regional scale; in other cases, local factors have dominated. The result has been a considerable diversity in form and process over the area.

Numerous new findings were made in the areas selected for detailed investigation over the course of this project (Chapters 5 through 13). The large variety of cause-and-effect relationships was far greater than originally anticipated at the onset of the study, primarily because, on a regional scale, the Edwards Plateau appears to be quite simple with regard to topographic, geologic, and hydrologic settings. Indeed, in comparison to many other karst regions in North America and on other continents, the Edwards Plateau karst (1) is seemingly underlain by rocks of rather uniform lithic character, (2) has a relatively uncomplicated structural setting consisting of nearly flat-lying strata which have been fractured systematically, and (3) has developed in response to progressive dissection of an extensive and continuous upland surface.

However, closer scrutiny within specific areas of karstic development has shown that some factors governing subsurficial processes have been dominant in some locales but have been suppressed in others. Moreover, superposition of factors and processes occurred to varying degrees over the plateau. In some isolated cases, fortuitous combinations of controlling factors and synergetic processes have produced superlative features of karst.

Only eight areas of karst developed on carbonate rock and an area of pseudokarst developed on granite were chosen for investigation under the constraints (primarily of time) of this study. The region-wide multiformity of karstic features and variegation of processes that formed them suggest that perhaps other, different configurations and genetic histories may occur but remain undiscovered at this writing.

SUGGESTIONS FOR FUTURE RESEARCH

> It is better to know some of the questions than all of the answers.
>
> —James Thurber
>
> The value of what one knows is doubled if one confesses to not knowing what one does not know. What one knows is then raised beyond the suspicion to which it is exposed when one claims to know what one does not know.
>
> —Arthur Schopenhauer

Additional Localities

The underpinnings of conclusions drawn from this study lie chiefly in the interpretation of phenomena within eight principal karstic areas of the Edwards Plateau. A primary suggestion for additional work would be the selection of additional sites over the region for analysis similar to that presented in Chapters 5 through 13. Supplementary localities should be widely dispersed over the plateau and include known areas of intensive development of caves. Suggested additional localities are provided in Figure 171 and Table 12. Note that the area of study should be extended somewhat to the west to include little-frequented areas of karst on the Stockton Plateau, where recent explorations have confirmed a great potential for horizontally and vertically extensive caves (e.g. Sorcerer's [Adam's] Cave, Terrell County).

Regionwide Geologic Studies

In the course of evaluating controls on speleogenesis within the selected areas, it became evident that where various aspects of the geologic setting differ from site to site, so do characteristic, morphologic features of the caves. Some regional trends are described in Chapter 14. Because of the large areal scope of this study, correlation of morphology of caves with geologic conditions has relied heavily on the availability of previously published reports that describe the topography, geology of bedrock, and hydrogeology of the region and the chosen areas. The influences that components of the physical setting have had on speleogenesis have been proposed. These results should now be tested in additional selected areas on the plateau.

More importantly, specific and pertinent factors of the physical setting should be selected and studied in greater detail in context with development of caves over the entire region. Measurements and determinations made from maps, in the field, and in the laboratory should be made uniformly and under controlled conditions for most known, cavernous areas. Some suggested topics for such consideration are as follows:

Lithostratigraphic Studies

Detailed study of the lithic character of cavernous sequences of rock and precise mapping of the stratigraphic position of passages in caves on a regionwide basis should establish the lithic properties that promote or retard dissolutional excavation. Standard petrographic and geochemical analyses of samples from both cavernous and non-cavernous beds should be employed. Quantitative data on composition (e.g. percent calcite or dolomite, insoluble residues), texture (e.g. grain-size, primary porosity), and diagenetic factors (e.g. cementation, induration, mineralogic changes) could then be related to the amount of dissolutional enlargement attained along particular stratigraphic horizons (provided other factors remain relatively constant

Table 12: Suggested Areas for Further Detailed Investigations[a]

A. Dryden Area, Terrell County:
The two deepest known caves in Texas, Sorcerer's Cave (170 m deep) and Wizard Well (118 m deep), lie 35 km west of the present area of study, in the vicintiy of Dryden on the Stockton Plateau. The lowest level of Sorcerer's Cave consists of a lengthy conduit containing an active stream. THe cave is also one of the longest in Texas (2690 m). These Caves have developed within Formations of the Edwards Group.

B. Big Lake Area, Reagan and Crockett Counties:
This, the northwestermost subarea of the karsted Edwards Plateau, includes Water Cave (approximately 370 m long) and Abyss Sinkhold (110 m and 34 m deep). These caves lie along the northern margin of deep topographic dissection and are developed in formations of the Edwards Group.

C. Pandale Area, Crockett and Val Verde Counties:
Several long caves lie within the deeply dissected terrain along the Pecos River, including Marshall Bat Cave (335 m long and 27 m deep). They have developed in formations of the Edwards Group.

D. Suggs Area, Crockett County:
The Suggs area includes O-9 Water Well, a major cave developed within the Edwards Group. It is 1370+ m long and 101 m deep and lies beneath a slightly dissected upland drained by tributaries of the Pecos, Devils, and Concho Rivers. The cave consists of a sizeable conduit containing an active stream.

E. Devils River Area, Val Verde County:
This area lies within the deeply dissected terrain of the Devils River drainage in central Val Verde County. Fern Cave, 1040 m long and 36 m deep, is one of the largest of the southwestern Edwards Plateau. It has developed within carbonate units of the Edwards Group.

F. Del Rio Area, Val Verde County:
Several long caves have developed within formations of the Edwards Group in the vicinity of the Rio Grande Valley. Diable Cave, a 2060 m-long conduit containing a stream, lies within a few meters of baselevel at the Rio Grande, just downstream from the Amistad Reservoir Dam.

G. Middle Concho River Area, Tom Green and Irion Counties:
This northernmost area of the Edwards Plateau includes Arden Cave, a 360 m-long cave along the Middle Concho River 30 km west of San Angelo. The cave has formed within units of the Edwards Group.

H. Eldorado Area, Schleicher County:
This area lies within the moderately dissected drainage basin of the South Concho River. Cave Y (610 m long and 32 m deep) has developed within units of the Edwards Group.

I. Sonora Area, Sutton, Val Verde, and Edwards Counties:
This is a southern and eastern extension of the Sonora area of Plate 1. In addition to the caves investigated in Chapter 6, the area includes Alma's Cave (335 m long), Word Cave (457 m long), and Harrison Cave (1830+ m long and 63+ m deep) in Sutton County and others in northern Edwards County. These caves lie within formations of the Edwards Group.

J. Carta Valley Area, Val Verde and Edwards Counties:
This is a major area of development of caves within the dissected southwestern Edwards Plateau. It includes Fawcett Cave (820 m long and 34 m deep) and H. T. Meirs Cave (779 m long and 100+ m deep) in eastern Val Verde County and Deep Cave (1180+ m long nad 91+ m deep), Blowhole (100+ m deep), Midnight Cave (900+ m long and 45+ m deep), Mighty Fine Pit (150+ m long and 59 m deep), and others in western Edwards Coutny. These caves have developed within formations of the Edwards Group.

K. Brackettville Area, Kinney County:
This area encompasses the westernmost, cavernous zone along the dissected Balcones Escarpment. Large Caves developed in formations of the Edwards Group include Webb Cave (900+ m long and 30 m deep), Green Cave (450+ m long), and Kickapoo cave (420+ m long and 40+ m deep).

L. Vance Area, Edwards and Real Counties:
The Vance Area includes the deeply dissected terrain along the headwaters of the East Nueces River. Caves developed in the Glen Rose Formation and units of the Edwards Group include Red Arrow Cave (300+ m long) and Devils Sinkhole (110 m long and 94 m deep).

M. Telegraph Area, Kimble, Edwards, and Kerr Counties:
This moderately dissected region along the South Llano River includes Fleming Bat Cave (360+ m long), and Schulze Cave (180+ m long and 37 m deep). The caves have developed in formations of the Edwards Group.

Table 12: continued

A. Uvalde Area, Uvalde County:
The Uvalde area consists of two distinct locations of large caves. The northern part of the area includes Indian Creek Cave (6710 m long and 40 m deep), the third longest cave in Texas. This cave, consisting of a well integrated, dendritic network of condiuts containing active streams, has formed within the Edwards Group. The southern part includes Rambie's Cave (900+ m long and 27 m deep) developed within the Salmon Peak Limestone.

B. Mason Area, Mason County:
This area includes the southwesternmost outcrop of Paleozoic rocks in the Llano Basin. Several caves of moderate length (60 to 120 m) occur within formations of the Ellenburger Group and within the Marble Falls Formation. Important caves include James River Ranch Cave No. 3, Mill Creek Cavern, and Zesch Ranch Cave.

C. Mountain Home-Leakey Area, Kerr, Real, and Bandera Counties:
This area encompasses the deeply dissected terrain along the headwaters of the Guadalupe, Frio, Sabinal, and Medina Rivers. Caves developed in formations of the Edwards Group include Stowers Cave (2390 m long and 38 m deep), Dan Auld Ranch Caves Nos. 1 and 2 (300+ and 600+ m long respectivesly), Cave of the Lakes (346 m long), Perry Water Cave (1060+ m long), and Station C Cave No. 1 (300+ m long).

D. Sabinal Area, Real, Bandera, Uvalde, and Medina Counties:
The Sabinal area lies along the dissected southern Balcones Escarpment. Caves developed in formaions of the Edwards Group include Frio Bat Cave (600+ m long and 30+ m deep), Santleben Cave (300+ m long), Blanket Creek Cave (170+ m long), Marguerite Cave (420+ m long and 40 m deep), and Valdina Farms Sinkhole (600+ m long and 46 m deep).

E. San Saba Area, McCulloch and San Saba Counties:
Caves of the San Saba area have developed in carbonate rocks of the Ellenburger Group and in the Marble Falls Formation. Most caves are small with few exceeding 200 m in length. Richland Springs Cavern (150 m long and 20 m deep) and Whiteface Cave (180+ m long and 61 m deep) are dignificant. This area represents the northernmost outcrop of karsted Paleozoic rocks in the Llano Basin.

F. San Antonio Arean, Bandera, Medina, and Bexar Counties:
This area includes much of the southeastern margin of the Edwards Plateau, along the Balcones Escarpment. Although most caves are small (less than 200 m in length), several are sizeable, including Wurzbach's Bat Cave (387 m long) and Robber Baron's Cave (1341 m long) developed in the Austin Chalk.

G. Bend Area, San Saba, Lampasas, Llano, and Burnet Counties:
This is an expansion of the Bend area of Plate 1. In addition to the caves investigated in Chapter 8, the area includes numerous small caves to the north, west, east, and south. All caves have developed within formations of the Ellenburger Group or within the Marble Falls Formation.

H. Johnson City Area, Blanco County:
The Johnson City area includes Buffalo Cave (1820+ m long), a sizeable cave consisting of long passages and sections of mazes. The cave has developed in the Cap Mountain Limestone.

I. Boerne Area, Kendall, Blanco, Bexar, and Comal Counties:
The Boerne area includes many of the long, well integrated caves of the Edwards Plateau. They contain perennial streams and are close to local baselevel in elevation. Most passages have developed in the lower Glen Rose Formation. SIgnificant caves include Honey Creek cave (the second longest [9250+ m] known in Texas), Cave Without a Name (4310 m long), Prassell Ranch Cave (2618 m long), Fair Hole (600+ m long and 26 m deep), Alzafar Cave (900+ m long), Cascade Caverns (400+ m long and 40+ m deep), Cascade Caverns Sinkhole (900 m long and 40+ m deep), Schwarz Cave (300+ m long), Spring Creek Cave (2400+ long), and Dead Deer Cave (300+ m long and 35+ m deep).

J. Smithwick Area, Burnet, Williamson, Blanco, and Travis Counties:
The Smithwick area includes several small caves (less than 400 m in length) developed int he Marble Falls Formation at the extreme eastern end fo the Paleozoic outcrop in the Llano Basin. The Caves lie close to the baselevel of the Colorado River.

K. Killeen Area, Lampasas, Coryell, and Bell Counties:
The Kileen area includes small caves developed wihtin the Glen Rose Formation and units of the Edwards Group of southern Coryell County and northern Bell County. Generally less than 200 m in length, the caves represent the northeasternmost cavernous zone of the Edwards Plateau.

L. Georgetown Area, Bell and Williamson Counties:
This is a southern and northern extension of the Georgetown area of Plate 1. In addition to the caves investigated in Chapter 12, the area includes many small caves (less than 300 m in length) and some moderately sized caves such as Hill's Cave (270+ m long), Beck Ranch Cave (670+ m long), Beck Sewer Cave (300+ m long), Cobb Caverns (760+m long), and steam Cave (600+ m long). These caves have developed within formations of the Edwards Group.

[a]Locations of these areas are shown in Figure 171. This list is an extension of areas listed in Table 8 and shown in Plate 1.

in the comparisons). Obviously, such a program of sampling and work in the laboratory would be quite extensive in time and effort, but the results would give a better understanding not only of speleogenesis in the region, but also of the diagenetic history of the carbonate rocks and of the evolution of principal, karstic aquifers

Structural Studies

This study has established that the morphology and orientation of chambers and passages in caves throughout the plateau are strongly guided by geologic structure. Suggestions for future work are two-fold. Primarily, the structural setting of any additional sites must be closely scrutinized to determine the effects of attitude and fractures (faults, joints, bedding-plane partings) on patterns of flow of the groundwater and on dissolutional excavation.Secondly, structural conditions should be more completely ascertained on a regional scale. The program of mapping fractures initiated along the southern margin of the plateau by Wermund and Cepeda (1977a,b), Wermund and others (1978) and Leonard (1977a,b; 1978) should be extended to the north in order to encompass the entire region of karst. Maps showing the distribution, extent, and orientation of fractures should be compared to known structural elements to identify and describe tectonic stresses responsible for fracturing. This interpretation may then be used in conjunction with data on attitude of beds to evaluate regional trends in the circulation of groundwater at the time of development of karst. This would help to explain the location and orientation of solutional conduits.

Additional studies on the role of faults in augmenting or impeding flow of groundwater would be of considerable interest. Supplementary examples of development of caves within zones of faults would aid in characterizing components of the evolving, carbonate aquifers, including places of recharge and discharge, compartments bounded by faults, avenues of enhanced transmissivity, and barriers imposed on migration of water

Hydrodynamic and Geochemical Studies

Few areas of the Edwards Plateau have a sufficient number of sinking streams or caves with accessible perennial streams for tracing groundwater, delineating flowpaths, determining the dynamic character of recharge and discharge, or monitoring the chemistry of water. However, where these studies are possible, it would be valuable to ascertain routes of drainage and to record rates of flow and geochemical parameters. Hydrographs of discharge and geochemistry would illustrate the response of particular karstic systems of flow to seasonal or short-term changes in input. The hydrodynamic character of present-day systems of flow should provide some insight regarding the circulation of groundwater during development of well integrated, but now abandoned, cave systems.

Primary sites for hydrodynamic and geochemical monitoring include the Powell-Neel Cave System of Menard County, the Indian Creek Cave System of Uvalde County, master conduits of the Bend area in San Saba County, numerous long caves containing active streams at or near baselevel in the drainage basins of the Guadalupe River and Cibolo Creek in Kendall, Comal, and Bexar Counties (e.g. Cave Without a Name, Alzafar Cave, Spring Creek Cave, Prassell Ranch Cave, Cascade Caverns System, Fair Hole, Honey Creek Cave, and others), 0-9 Well in Crockett County, and Sorcerer's (Adam's) Cave in Terrell County.

Further investigations are needed of paleoflooding in caves and its role on speleogenesis on the plateau. The principal efforts would address (1) recognition and careful mapping of flood-developed morphologic features in caves (e.g. variable cross-sections of passages, joint-spurs, pockets in ceilings,redissolved

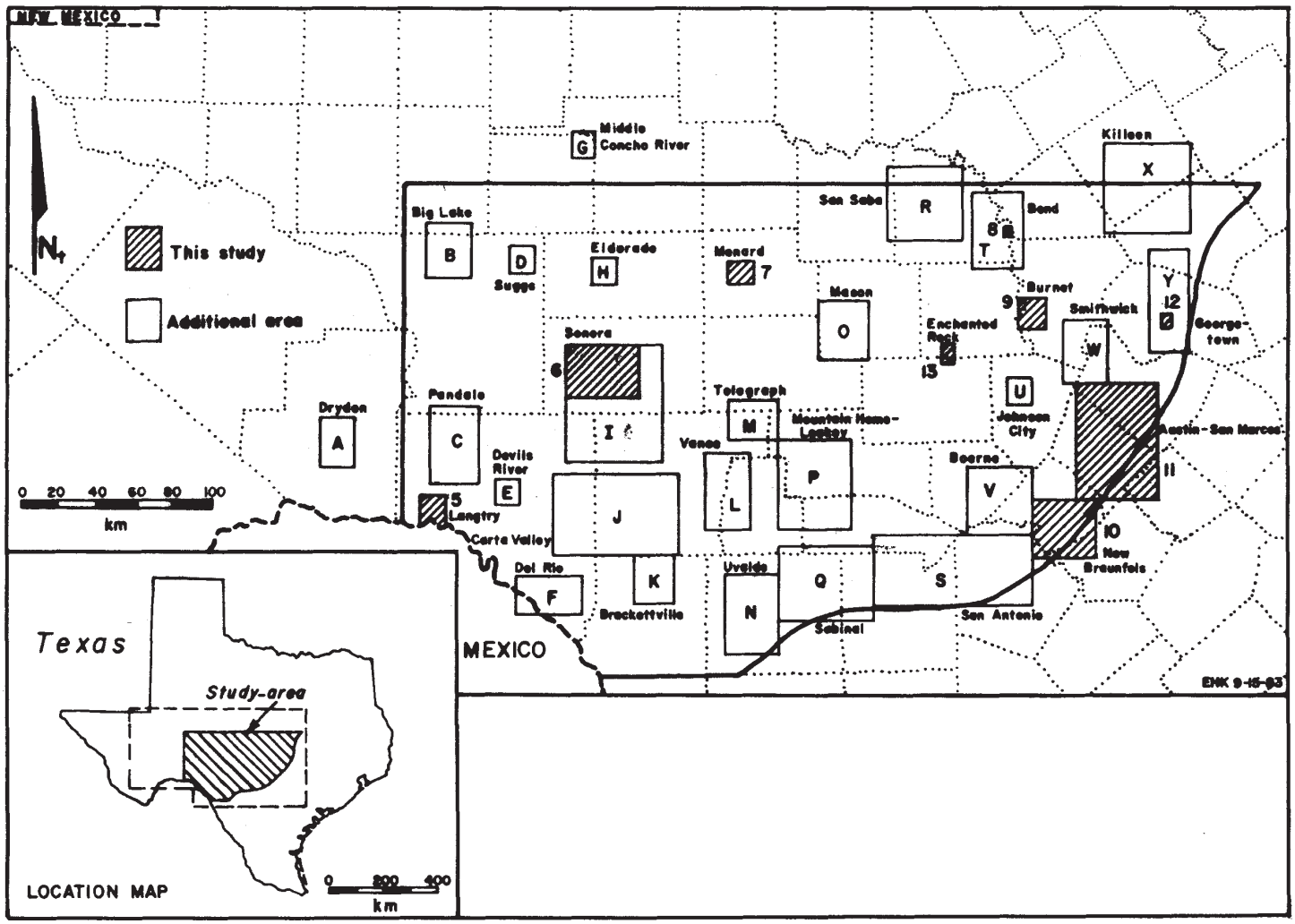


Figure 171: Suggested areas for further detailed investigations (shown dotted). Letters correspond to lettered entries on Table 12. Shaded areas are those of this study. Numbers correspond to those of appropriate chapters.

speleothems, etc.), (2) correlation of these features with any constrictions or blockages along avenues of flow, and (3) establishment of a relative chronology among the episodes of modification by floodwater in context with changes within the caves (e.g. breakdown, blockage, diversion, etc.) and with long-term, climatic changes (i.e. alternation between pluvial and warm climatic conditions)

Quantitative Geomorphic Studies

Speleogenesis is intimately related to the evolution of the surficial topography of the Edwards Plateau, including such processes as fluvial erosion and regional denudation of the uppermost, exposed strata. One key aspect of formulating a unified chronology of speleogenesis over the region would be the establishment of a history of sequential changes in baselevel brought about by the incision of major streams into the uplifted plateau. Quantitative analysis of geomorphic data obtained from maps, remotely-sensed imagery, and field investigations, should aid considerably in relating development of karst to surficial processes. Some recommended lines of research include (1) morphometric analysis of drainage-composition, (2) hypsometric analysis of selected drainage-basins, and (3) analysis of topographic slopes, stream incision, and widening of valleys. These studies should contribute to a better understanding of the geomorphic evolution of the plateau by providing information on rates of adjustment of streams and identifying areas of anomalous behavior (e.g. stream-capture, abrupt changes in slopes or drainage-composition, etc.).

Regional Synthesis and Characterization

Once regional geologic studies have been completed in sufficient detail, it would then be possible to develop a unified hydrogeomorphic history of the karst of the Edwards Plateau. Upon the formulation of a model for the development of the karst it would be possible to (1) characterize elements of the karstic landscape for comparison with other well studied regions of karst, (2) incorporate the evolutionary development of the plateau into the geologic history of central Texas, (3) integrate the findings with the current knowledge of Quaternary events in North America, and (4) compare and contrast the karst of the Edwards Plateau with other karstic areas worldwide.

Contributions to Knowledge of Speleogenesis

Conditions and processes of karstification that are unique to the Edwards Plateau and those that are similar to phenomena of other regions should be identified. In this way, attributes from this large, newly studies region may be added to the knowledge already in hand for many other significant karstic regions. The result would contribute to a better understanding of speleogenetic processes.

Applications Toward Resources and Environmental Problems

Knowledge gained from the study of the karst of the Edwards Plateau ma be useful in future management of the region as it becomes more populous and developed. An effort should be made to apply the knowledge derived form this and subsequent studies toward effective evalutation fo the region's resourves and potentional environmental problems. These include availability and quality fo water, stability of slopes, carrying capacity of the land and land-use, and preservation of the aesthetic integrity of the region.

SUMMARY

The culmination of this initial phase in the investigation of the karst of the Edwards Plateau has given rise to divergent perspectives. Despite advances made toward an understanding of the hydrogeomorphic evolution of the karst, the study has exposed a plethora of new questions which will eventually need to be addressed. The generalized, regional synthesis of development of karst presented herein encompassses numerous examples from several disparate settings and should serve well as a first-order model for the Edwards Plateau. However, in all likelihood, this model will need to be refined as investigations continue, particularly at additional sites where governing conditions may have differed from those encountered in this investigation. Moreover, global advances in the fields of geomorphology and hydrogeolgoy of karst will also dictate a re-evaluation of the Edwards Plateau at some later date. In this context, the findings of this study should provide a substantial basis for future research.

If you are out to describe the truth, leave elegance to the tailor.
—Albert Einstein

All elevated thinking ends in a sigh.
—Paul Valery

APPENDICES

APPENDIX A
DEFINITIONS OF KARST AND PSEUDOKARST

Geomorphologists have differed in their definitions of the terms "karst" and "pseudokarst" (Quinlan, 1966, 1978). Several definitions are listed here to demonstrate the variability in meaning among these.

DEFINITIONS OF KARST

1. "(Karst is) a terrain, generally underlain by limestone, in which the topography is chiefly formed by the dissolving of rock, and which is commonly characterized by Karren, closed depressions, subterranean drainage and caves." (Monroe, 1970, p. K11).

2. "(Karst is) a solution-controlled landform type, characterized by an exclusive surface morphology, subsurface drainage and collapse features, which is specifically developed in calcareous rock. These elements remain constant even though climate, structure, and petrography generate greatly diverse-surface features." (Addison, 1973, p. 333).

3. "Karst signifies *terrain with distinctive characteristics of relief and drainage arising primarily from a higher degree of rock solubility in natural waters than is found elsewhere.* The word is also used adjectively to refer to rock, water, streams, caves and other features making up such landscape" (Jennings, 1971, p. 1).

4. "(A karst region is a) region underlain by compact and soluble rocks in which appear distinctive surficial and subterranean features, caused by solutional erosion. *Karst regions are characterized by limestones and other soluble rocks at or near land surface that have been modified by solutional erosion. (Note: the term can also be applied to any region made up of soluble rocks: gypsum, salt, etc.)*" (Paloc, 1975, p. 159).

5. "(Karst is) land where the underground karstic drainage due to fissured rock and an efficient solution occurs and where specific surface and subsurface features can develop. Only the presence of the karstic drainage is needed for the denomination 'karst'." (Gams, 1973, p. 13; 1974, p. 330; translated in part by Quinlan, 1978).

6. "(Karst is) all limestone regions where characteristic landforms are found. *Karstification* (is) the action of acid-charged infiltration waters on the calcium carbonate of limestones, the dissolution being made irregular by structure and being associated with processes of mechanical erosion and collapse phenomena.... By extension the same term is applied to the evolution of a limestone terrane having an internal and external morphology more or less like that of the Yugoslav karst. The *karstification* is equally suitable for 'pseudokarstic' rocks (sandstone, gypsum, salt, granite, etc.) when one finds certain relief forms of karst: karren, dolines, caves, pits, etc." (Fenelon, 1968, p. 48 and 49; translated in part by Quinlan, 1978).

7. "(Karst is) originally the proper name of the mountain landscape in the vicinity of Trieste, now the designation for all morphologically- and hydrologically similar regions in which underground drainage, as a result of jointing and solubility of the rock, and in which characteristic surface and subsurface karst phenomena, due to corrosion, can be found" (Trimmel, 1965, p. 57-58, translated in part by Quinlan, 1978).

"Karst is the process of chemical and partly mechanical action of non-fluvial surface and subsurface waters on soluble permeable rocks (carbonates, gypsum and anhydrite, salt, sodium carbonate). The effects are not only surficial and underground degradational features but also aggradational features formed by precipitation from solution and by collapse. The formation and destruction of underground cavities may be accompanied by subsidence, collapse, and even by local earthquakes. In a broad sense, karst is the forms themselves as well as the process of their formation" (Maksimovich, 1963, p. 7; translated in part by Quinlan, 1978).

9. "The term *karst* unites specific morphological and hydrological features in soluble (mostly carbonate) rocks. Morphological features include karren, dolinas, jamas, ponors, uvalas, poljes, caves, caverns, etc. Hydrological features include basins of closed drainage, lost rivers, estavelles, vauclusian springs, submarine canyons, more or less individualized underground streams and incongruity of surface and underground divides. *Karst* is understood to be the result of natural processes in and on the earth's crust caused by solution and leaching of limestones, dolomites, gypsum, halite, and other soluble rocks" (Food and Agriculture Organization, 1972, p. 21).

10. "(Karst is a) type of topography that is formed on limestone, dolomite, gypsum, and other rocks by dissolving or solution, and that is characterized by closed depressions or sinkholes, caves, and underground drainage" (Bates and Jackson, 1980, p. 337, as modified from Gary and others, 1972, p. 383).

DEFINITIONS OF PSEUDOKARST

1. "Pseudokarst is karstlike terrain produced by a process other than the dissolving of rock, such as the rough surface above a lava field, where the ceilings of lava tubes have collapsed. Features of pseudokarst include lava tubes, lava stalactites, and lava stalagmites" (Monroe, 1970, p. K14).

2. "Pseudo-karst is the name given to phenomena developed on non-limestone rocks, but which resemble karstic forms.... Pseudo-karst phenomena have been defined as karst-like features in non-carbonate and non-evaporite rocks which have the morphology (though rarely the hydrological flow pattern) of karst features produced in carbonate rocks. In general, such pseudo-karstic features are restricted to the superficial and smaller landforms" (Sweeting, 1973, p. 306).

3. "A pseudokarst region is a region presenting forms similar to those of karst in which the rocks are not soluble. In a broader sense, the term is utilized to designate any karstlike terrain produced by a process other than the dissolving of rocks" (Paloc, 1975, p. 159).

4. "A pseudokarst terrain is a terrain with features similar to karst but formed in non-soluble rocks, as by melting of permafrost or ground ice, collapse after mining, and by outflow of liquid lava from beneath its solidified crust" (Food and Agriculture Organization, 1972, p. 32).

5. "(Psuedokarst includes) only thermokarst, piping, and forms created predominantly by these processes..." (Otvos, 1976, p. 1026).

6. "(Pseudokarst is a) topography that resembles karst but that is not formed by the dissolution of limestone, usually a rough-surfaced lava field in which ceilings of lava tubes have collapsed. Such an area is characterized by tunnels, tubes, stalactites, and stalagmites of lava" (Gary and others, 1972, p. 574).

7. "(Pseudokarst is a) topography that resembles *karst* but was not formed by the dissolution of rock. Processes and forms involving *piping* and *thermokarst* are included (Otvos, 1976); some authors also include terrain characterized by lava tubes, sea caves, and blowouts. Pseudokarst has been applied to covered karst, and to karst produced by the dissolution of rocks that are relatively insoluble, such as quartzite and granite, but more general usage regards these as varieties of true karst" (Bates and Jackson, 1980, p. 506).

8. "(Pseudokarst is) an assemblage of landforms resembling karst but formed by processes where the dominant removal of material is not in solution. Volcanokarst, talus caves, and piping phenomena...(are) prime examples" (Beck, 1980, p. 25).

9. "Pseudokarst phenomena are landforms, usually small in scale, that are similar in appearance to those created by karstification (chemical weathering, usually dissolution), yet in actuality are formed by other processes" (Kastning, 1980c, p. 25).

The definitions of karst and pseudokarst used in this dissertation are those proposed by Quinlan (1978). They are quoted in Chapter 1.

APPENDIX B
GEOLOGICAL TABLES

STRATIGRAPHIC UNITS OF THE EDWARDS PLATEAU

The stratigraphic nomenclature used in this investigation follows that of the Geologic Atlas of Texas published at a scale of 1:250,000 (Barnes 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b). The karstic phenomena studied in this dissertation have developed within two primary stratigraphic sequences: (1) the Gorman Formation of the Ellenburger Group (Ordovician) and (2) numerous formations of the Cretaceous System. Lithostratigraphic information about the former is readily available from a few key references (most notably Cloud and Barnes, 1946a) and from summaries of previous geologic work presented in Chapters 8 and 9. However, similar data about the Cretaceous rocks is widely disbursed in the literature. For convenience and reference during the course of this investigation descriptive information on Mesozoic and Cenozoic rocks, including type of rock, lithologic characteristics, thickness, occurrence, and significance to speleologic problems, has been assimilated into tabular form (Table 13).

GEOLOGIC COLUMN OF PRE-CRETACEOUS ROCKS (PLATE 4)

Rocks exposed in the Llano Basin are largely Precambrian and Paleozoic. Plate 4 is a geologic column of all Pre-Cretaceous units shown on sheets of the Geologic Atlas of Texas (Barnes, 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b) that crop out within the area of study. No vertical scale is implied on the plate. Lithologic and stratigraphic information shown thereon is compiled from Barnes (1954, 1976a, 1981a), Barnes and Bell (1977), Barnes, Cloud, and Warren (1947), Barnes and others (1945, 1953, 1966), Bell and Barnes (1961, 1972), Bridge and others (1947), Cheney (1940), Clabaugh and McGehee (1972), Cloud and Barnes (1946a, 1957), Cloud and others (1945, 1957), Oetking and Feray (1963), Plummer (1943), Plummer and Moore (1921), Renfro and others (1973), and Sellards (1931, 1932a).

CORRELATION CHART OF CRETACEOUS ROCKS (PLATE 5)

Stratigraphic units of the Cretaceous System of central Texas are shown in the correlation chart of Plate 5. The four individual columns correspond to regional paleogeographic and structural elements shown in Figure 9. No vertical scale is implied on the plate. Stratigraphic nomenclature is also that of the Geologic Atlas of Texas. Descriptive information on each unit on the chart is presented in Table 13. The chart is compiled from Adkins (1932), Adkins and Lozo (1951), Barnes (1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b), Bishop (1967), Boone (1968), Cuyler (1929), Darton and others (1937), Fisher and Rodda (1966, 1967a, 1969), Forgotson (1957), Freeman (1961), Hayward and Brown (1967), Lozo (1959), Lozo and Smith (1964), Lozo and Stricklin (1956), Martin (1961, 1967), Moore (1959, 1961a,b, 1964, 1967) Moore and Martin (1966), Oetking and Feray (1963), Pessagno (1969), Renfro and others (1973), Rives (1962), Rodda and others (1966), Rose (1968, 1970, 1971, 1972), Shelburne (1959), Stephenson (1927, 1929, 1937), Stephenson and others (1942), Stricklin and Smith (1973), Stricklin and others (1971), Wilbert (1963, 1967), and Young (1959, 1963, 1965, 1967, 1972a,b).

GEOLOGIC MAP OF THE EDWARDS PLATEAU (PLATE 3)

The nine sheets of the Geologic Atlas of Texas that cover the area of study (Barnes, 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b) were used as the geologic base for most of this study. Where available, geologic maps of larger scale were used in the areas selected for detailed investigations (Chapters 5 through 13). A simplified geologic map of the area of study (Plate 3) is included herein for completeness and for comparison with other region-wide maps showing physiographic divisions (Figure 2), surficial drainage (Plate 2), geologic structure (Figure 10), soils (Figure 6), aquifers (Figures 4 and 5), and distribution of caves (Plate 1).

STRUCTURAL MAP OF CENTRAL TEXAS (FIGURE 10)

The following sources were consulted in compiling the map of structural and tectonic elements of central Texas (Figure 10): Barnes (1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b), Caran and others (1981), Cohee (1961), Eardley (1951), Eubanks (1964), Flawn and others (1961), Fowler (1956, 1957), Freeman (1968), Nicholas and Rozendal (1975), Oetking and Feray (1963), Rall and Rall (1958a), Renfro and others (1973), Sellards (1934), Sellards and Hendricks (1946), Van der Gracht (1931a,b), Vertrees and others (1959), Walper (1977), Webster (1978, 1980, 1982), Weeks (1945b), Wood and Walper (1977), Woodruff (1980), and Young (1960). Detailed maps at larger scale are available for various localities. Those used in the investigations of selected areas of the Edwards Plateau are cited in the appropriate places in Part II (Chapter 5 through 13).

Table 13: Mesozoic and Cenozoic Rocks in Central Texas

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Alluvium (Holocane)	Unconsolidated	Floodplain and low terrace deposits; cobbles, gravel, sand, silt, clay, and organic matter; silt and clay, generally calcereous; sand, largely quartz; cobbles and gravel, siliceous, mostly chert and limestone, along Colorado River and Rio Grande includes igneous and sedimentary rocks from central and Trans-Pecos Texas, Mexico, and New Mexico.	Up to 11 m	Low terrace deposits near floodplane level, locally in stream channels throughout area.	Possible correlation with cave deposits and some passage levels.	
Windblown sand (Holocane)	Unconsolidated	Sand and silt in sheets, local dunes.	Less than 1 m	Minor occurrence in Reagan County.		
Low terrace (Holocene)	Unconsolidated	Some alluvium; gravel, mud, silt, clay, and organic material.		Mostly low terrace deposits above flood level along entrenched streams, southeast Edwards Plateau; principally along parts of Guadalupe and Media Rivers and Cibolo Creek.	Indicates former floodplains.	
Quaternary deposits (Pleistocene-Holocene)	Unconsolidated	Intimately associated alluvium and boulders, cobbles, and pebbles of Cretaceous limestone and chert; caliche overlain by brown silt in gently sloping areas.	Undetermined	In low lying areas in northwestern part of area.		
Caliche and gravel deposits (Pleistocene-Holocene)	Unconsolidated	Caliche and gravela, mostly covered by soil.	Up to 5 m	Forms partly alluciated flats to stream valleys, above level of fluvistile terraces, partly stream out, in northwestern part of study area.		

TABLE CONTINUED ON FOLLOING PAGES

[a]Compiled principally from Geologic Atlas of Texas (Barnes 1970, 1974a,b,c, 1975, 1976a, 1977, 1981a,b) with additions from primary references cited in this table.
[b]Units of bedrock listed in approximate stratigraphic sequence with youngest units at top of table and oldest at bottom. Stratigraphic nomenclature is that of the Geologic Atlas of Texas
[c] Unit not mapped separately on sheets of the Geologic Atlas of Texas (scale 1:250,000).

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Fluvial terrace deposits (Pleistocene-Holocene)	Unconsolidated	Gravel, sand, silt, and clay, commonly with pebbles and cobbles of chert, quartzite, igneous rock, metamorphic rock, caliche in northwest; gravel along Guadalupe River, siliceous, coarse, along Colorado River mostly limestone, dolomite, chert, quartz; sand, mostly quartz; adjacent to Edwards Plateau, predominantly gravel, limestone, and chert.		Along streams throughout area.	Terraces may correspond to variations in climate during Pleistocene	Baker, 1977; Baker and Penteado-Orellara, 1977, 1978.
Onion Creek Marl (Pleistocene)	Unconsolidated	Sand, clay, and gravel	Up to 15 m	Broad terraces in the vicinity of Buda, Hays County, and downstream along Onion Creek. Possibly a correlate of the Leona Formation.	May correspond to a former drainage system on the Edwards Plateau.	DeCook, 1960, 1963.
Leona Formation (Pleistocene)	Unconsolidated	Sand, clay, and gravel with red and reddish gray silt.	Up to 15 m	Forms broad terrace southeast of Kyle, Hays County. Upper surface 15-17 m above flood-plain level. Possibly a correlative of the Onion Creek Marl.	May correspond to a former drainage system on the Edwards Plateau.	Plummer, 1932
High gravel Deposits (Pleistocene)	Unconsolidated	Upper silty clay unit; lower coarse gravel unit.	Up to 15 m	In vicinity of Austin and along eastern boundary of area between Austin and Georgetown. Possibly correlates with Onion Creek Marl.	Corresponds to former drainage system on the Edwards Plateau.	
Playa deposits (Pleistocene)	Unconsolidated	Clay and silty, sandy, light gray, usually covered by thin deposit of recent sediment (Wisconsinan).		In shallow depressions; a few deposits north of Del Rio; common in northwestern part of area.	Deposits are in karstic depressions.	Pool, 1976, 1977.
Tahoka Formation (Pleistocene)	Weakly consolidated sediments	Lacustrine clay, silt, sand, and gravel, locally calcareous, selenitic. Clay and silt, sandy, indistinctly bedded to massive, weakly coherent, various shades of light gray and bluish gray. Sand, fine- to coarse-grained quartz, indistinctly bedded to massive, friable, gray, grades to gravel at margins of deposits.	Up to 8 m	Single deposit less than 4 km in extent at Big Lake, Reagan County.		Evans and Meade, 1945
Alluvial fan deposits and colluvium (Pleistocene)	Caliche-cemented sediments.	Poorly sorted, angular to rounded rock fragments of locally derived material.		Common in lowlands north of Balcones Escarpment in eastern Val Verde, Kinney, and western Uvalde Counties.	Related to deep dissection.	

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Willis Forma tion (Pleistocene)	Unconsolidated	Gravel, sand, silt, and clay; mostly gravel in area; fluviatile.	Less than 30 m	Small patches southeast of Balcones Escarpment between New Braunfels and San Marcos. May correlate with Uvalde Gravel.	Corresponds to former drainage sysem on the Edwards Plateau.	Weeks, 1945a; Forney, 1950; Doering, 1956
Willis Forma tion (Pleistocene)	Unconsolidated	Gravel, sand, silt, and clay; mostly gravel in area; fluviatile.	Less than 30 m	Small patches southeast of Balcones Escarpment between New Braunfels and San Marcos May correlate with Uvalde Gravel.	Corresponds to former drainage system on the Edwards Plateau.	Weeks, 1945a; Forney, 1950; Doering, 1956
Uvalde Gravel (Pliocene-Pleistocene)	Caliche cemented gravel.	Gravel, some boulders up to 0.5 m in diameter; well rounded cobbles of chert, some cobbles of quartz, limestone, and igneous rock.	1 to 10 m	Occupies topographically high areas not associated with present drainage; forms extensive deposits in Medina, Uvalde, Kinney, and Val Verde Counties south of Balcones Escarpment. May correlate with Willis Formation.	Corresponds to former drainage system on the Edwards Plateau	Byrd, 1971.
Indio Formation (Eocene)	Sandstone, shale, siltstone, and lignite	Sandstone, fine-grained, thin bedded to laminar, varicolored in grays, yellows, greens, and browns; shale, sandy, carbonaceous, greenish-gray to medium gray and chocolate brown; siltstone, thin-bedded to laminar; numerous calcareous, arenaceous, and ferruginous concretions.	75-135 m	South of Balcones Escarpment in Kinney and Uvalde Counties. Correlative with Wilcox Formation east of the Frio River.		Plummer, 1932.
Igneous Rocks (Upper Cretaceous)	Basalt	Alkalic basalt and pyroclastics altered to montronite.		Generally in form of sills, laccoliths, volcanic necks, and dikes; along and just south and east of Balcones Escarpment.		Lonsdale, 1927.
Kemp Clay (Upper Cretaceous)	Clay	Calcareous, locally silty, massive, thinly laminated, conchoidal fracture, medium dark gray, weathers medium gray.	25 to 60 m; thins locally	In northeastern part of area. Included in Navarro Group.		Adkins, 1932; Pessagno, 1969.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Escondido Formation (Upper Cretaceous)	Shale, siltstone, and sandstone	Shale, gray to bluish gray; siltstone, brownish yellow, thin-bedded; sandstone, fine-grained, calcareous, in part asphaltic, argillaceous, gray, flaggy bedded; some pyroclastic material.	60 to 150 m	Generally poorly exposed; south of Balcones Escarpment in Kinney, Uvalde, and western Bexar Counties. Included in Navarro Group, correlates with Kemp Clay and Corsicana Marl in northeastern part of area.		Adkins, 1932; Pessagno, 1969; Cooper, 1970.
Corsicana Marl[c] (Upper Cretaceous)	Marl	Calcareous, montmorillonitic claystone.		Along eastern edge of Balcones Escarpment; included in Navarro Group.		Adkins, 1932.
Olmos Formation (Upper Cretaceous)	Clay, sandstone, coal and fireclay	Irregularly stratified, interfingering lenses of sandstone and clay, ferrugenous concretions and silicified wood common.	Up to 150 m	Just southeast of area in northeast Maverick County. Feathers out eastwardly. Included in Navarro Group.		Adkins, 1932; Pessagno, 1969.
Nacatoch Sand (Upper Cretaceous)	Sandstone	Quartz, fine to medium-grained, glauconitic, poorly sorted, friable, greenish gray; local lenses of silty clay; calcite-cemented sandstone in thin hard beds in lower and upper parts.	Up to 55 m	Crops out east of area in northeastern part. Recorded in the subsurface. Feathers out southwardly. Absent south of San Gabrial River. Included in Navarro Group.		Adkins, 1932; Pessagno, 1969.
Neylandville Marl[c] (Upper Cretaceous)	Clay and marl	Dominantly montmorillonitic, silt-size quartz becomes more abundant upward, calcite fragments common, glauconitic, disseminated pyrite, blocky with conchoidal fracture, light medium gray, weathers light brown to light gray and becomes slightly fissile.	Up to 82 m	Crops out east of area in northeastern part. Recorded in the subsurface. Absent south of San Gabrial River. Commonly mapped with Marlbrook Marl as a unit. Included in Navarro Group.		Adkins, 1932; Pessagno, 1969.
Marlbrook Marl[c] (Upper Cretaceous)	Clay, marl, sandstone	Clay, calcareous, locally silty; marl and clay, glauconitic, contains concretions of limonite, siderite, and siliceous limestone, conchoidal fracture, medium dark gray, weathers medium gray; sandstone, fine-grained, and siltstone, yellow brown.	150 to 180 m	Along southern and eastern edge of Balcones Escarpment from Bexar to Travis County. Also known as "upper Taylor Marl" and Bergstrom Formation in Austin vicinity. Included in Taylor Group.		Young, 1965; Pessagno, 1969.
San Miguel Formation (Upper Cretaceous)	Sandstone, limestone and clay	Fossiliferous sand, sandy limestone, and unfossiiferous clay superficially like that in underlying Upson Clay, clay increases in amount upward	Up to 240 m	Small outcrops along southern boundary of Kinney County; equivalent to Marlbrook Marl to the east.		Adkins, 1932; Pessagno, 1969.
Upson Clay (Upper Cretaceous)	Clay	Slightly calcareous, dark-gray to greenish-gray; weathers yellowish-gray	Approximately 150 m	Small outcrops in southern Kinney County; feathers out eastwardly; included in Taylor Group		Adkins, 1932; Pessagno, 1969.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Anacacho Limestone (Upper Cretaceous)	Limestone and marl	Limestone, reefy, thick-bedded, in part crossbedded; light-yellow to yellow-brown and light-gray, yellow; in part sandy, some volcanic rock fragments and weathered, rusty bentonite beds.	Up to 150 m	Along southern boundary of study area; ends abruptly in south central Kinney County and thins to feather edge in western Bexar County. Included in Taylor Group.		Adkins, 1932.
Pecan Gap Chalk (Upper Cretaceous)	Chalk and chalky marl	More calcareous westward, very light yellow to yellowish brown, weathers to form moderately deep soil; chalk in lower part grading upward to chalky marl with microgranular calcite in clay matrix, gray, weathers light gray and white; grades laterally in places to marl.	30 to 120 m	Along the eastern edge of the study area from eastern Medina County to Bell County. Correlates with the Anacacho Limestone to west; feathers out northward. Included in Taylor Group.		Adkins, 1932; Ellisor and Teagle, 1934; Young, 1965; Pessagno, 1969.
Wolfe City Formation (Upper Cretaceous)	Marl, sand, sandstone, and clay	Marl, sandy, silty, dark gray to light gray and brown, interbedded with thin sandstone lenses cemented by sparry fine- to coarse-grained calcite, grain size increases northward; uncemented sand increases northward; clay, glauconitic, phosphate and hematite nodules, dark gray to brown.	Up to 90 m	Crops out northeast of study area. Recorded in subsurface. Feathers out to south. Included in Taylor Group.		Adkins, 1932; Pessagno, 1969.
Wolfe City Formation (Upper Cretaceous)	Marl, sand, sandstone, and clay	Marl, sandy, silty, dark gray to light gray and brown, interbedded with thin sandstone lenses cemented by sparry fine- to coarse-grained calcite, grain size increases northward; uncemented sand increases northward; clay, glauconitic, phosphate and hematite nodules, dark gray to brown.	Up to 90 m	Crops out northeast of study area. Recorded in subsurface. Feathers out to south. Included in Taylor Group.		Adkins, 1932; Pessagno, 1969.
Ozan Formation (Upper Cretaceous)	Clay	Marly, calcareous content decreases upward, montmorillonitic, some glauconite, phosphate pellets, and hematite and pyrite nodules, variable amounts of silt-sized quartz and calcite fragments become more abundant upward, blocky with conchoidal fracore abundant upward, blocky with conchoidal fracture, light gray to brown; weathers light gray to grayish orange and white, develops poor fissility.	150 to 180 m	Along eastern boundary of study area from Austin to Bell County; known also as “lower Taylor Marl” and Sprinkle Formation.		Young, 1965; Pessagno, 1969.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Burt Ranch Member[c] of the Segovia Formation (Lower Cretaceous)	Limestone	Marly and bioclastic; light yellowish gray, thin-bedded to massive, nodular, wispy, whole mollusc marly micrite; thin-bedded miliolid biomicrite and fossiliferous marl most common at base and top; lower marly interval is Dr. Burt zone.	18 to 20 m	Narrow outcrop at base of Segovia Formation. Included in Segovia Formation.		Rose, 1968, 1972.
Fort Terrett Formation (Lower Cretaceous)	Limestone and dolomite	Uppermost part, porcelaneous aphanitic limestone, collapse breccia (Kirschberg Evaporite), chert, and recrystallized limestone; in middle light to dark gray cherty dolomite, miliolid, shell fragment, rudistid limestone, and medium brownish gray dolomite; in lower part, nodular limestone (Basal Nodular Member), with thin, yellow clay at base.	35 to 90 m	Extensive outcrops in uplands north of Balcones Fault Zone and west of Llano Basin, along margin of Segovia Formation; thickens southward. Includes Kirschberg Evaporite, Dolomitic Member, Burrowed Member, and Basal Nodular Member. Included in Edwards Group.	Cavernous, karsted. Together with Segovia Formation and Trinity Group forms unconfined Edwards-Trinity aquifer.	Moore, 1967; Rose, 1968, 1969, 1972.
Kirschberg Evaporite[c] (Lower Cretaceous)	Limestone and gypsum	Crystalline limestone, collapse breccia, gypsum and porcellaneous limestone; pink to gray, medium- to coarse-crystalline, massive to thin-bedded, cherty, crystalline limestone and travertine containing intervals of limestone-, dolomite-, and chert-fragment collapse breccia, overlain by light gray, thin-bedded brittle porcellaneous to lithographic micrite, commonly collapsed and brecciated; in northern Gillespie County, interval contains massive gray gypsum.	15 m + or -	Exposed along outcrops of uppermost Fort Terrett Formation. Included in Fort Terrett Formation and Kainer Formation in vicinity of Balcones Fault Zone.	Solution-collapse structures, exposed gypsum is karsted.	Rose, 1968, 1969, 1972; Facundus, 1968; Moore and others, 1969; Facundus and Moore, 1970.
Dolomitic Member[c] of the Fort Terrett Formation (Lower Cretaceous)	Dolomite	Light to medium gray, thin-bedded to massive, fine- to medium-crystalline, dolomitized micrite and mollusc-fragment biomicrite; irregular chert nodules scattered throughout; scattered possible hard crusts, flat clasts, disrupted zones, calcite pseudomorphs after gypsum nodules; some thin beds of miliolid biosparite intimately interbedded with dolomite; rudist biostromes at base in some places.	15 m + or -	Exposed as outcrop band within Fort Terrett Formation. Included in Fort Terrett Formation.	Locally cavernous and karsted.	Rose, 1968, 1972.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Burrowed Member[c] of the Fort Terrett Formation (Lower Cretaceous)	Limestone	Light gray, massive, resistant, porous, buttowed nodular mollusc biomicrite; some even thin beds of miliolid and mollusc-fragment biosparite above, rippled and cross-bedded in part, few thin dolomite interbeds.	20 m + or -	Exposed as outcrop band within Fort Terrett Formation. Included in Fort Terrett Formation.	Cavernous, and karsted.	Rose, 1968, 1972.
Basal Nodular Member[c] of the Fort Terrett Formation (Lower Cretaceous)	Marl and limestone	Marl and marly limestone, light yellowish gray, soft, recessive, nodular oyster marl grading upward to whole mollusc biomicrite; sandy and silty in lower half in some places; lower boundary generally abrupt, in many places red stained, irregular, bored, plastered with oysters.	15 m + or -	Crops out as basal member of Fort Terrett Formation.		Rose, 1968, 1972.
Devils River Formation (Lower Cretaceous)	Limestone and dolomite.	Hard, miliolid, pellet, rudistid, shell fragment biosparite and lime mudstone; locally dolomitized, brecciated, and chert bearing; rudistid mounds are more common in upper part; nodular limestone in basal part.	210 m + or -	Exposed along Devils River Trend in south part of study area from Langtry, Val Verde County to eastern Medina County. Included in Edwards Group.	Cavernous, karsted; a component of the Edwards aquifer.	Rose, 1968, 1972; Lozo and Smith, 1964.
Salmon Peak Formation (Lower Cretaceous)	Limestone and mudstone	Consists of Upper Grainstone Unit and Pryor Member (see below).	95 to 115 m	Outcrop along 30 km band south of Devils River Trend from Del Rio to Uvalde. Included in Edwards Group.	Locally cavernous.	Lozo and Smith, 1964.
Upper Grainstone Unit[c] of the Salmon Peak Formation (Lower Cretaceous)	Limestone	Granular, abundant caprinid and other shell fragments, cross-bedded.	23 m	South of Devils River Trend from Del Rio to Uvalde. Included in Salmon Peak Formation.	Locally cavernous.	Lozo and Smith, 1964.
Pryor Member[c] of the Salmon Peak Formation (Lower Cretaceous)	Mudstone	Abundant large chert masses, white.	70 to 90 m	South of Devils River Trend from Del Rio to Uvalde. Included in Salmon Peak Formation.	Locally cavernous.	Lozo and Smith, 1964.
McKnight ′ Formation (Lower Cretaceous)	Limestone and shale	Upper third, lime mudstone, thin-bedded, locally contains chert layers, solution zones, and collapse breccia; intermediate unit, argillaceous lime mudstone, laminated, fissile, black; lower half limestone, granular, thin chert layers, abundant shell fragments and pellets, overlain by lime mudstone with solution zone and collapse breccia.	45 m	Northeastern Kinney County and western Uvalde County. Included in Edwards Group.	Locally cavernous, solution-collapse zones.	Lozo and Smith, 1964.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Austin Chalk (Upper Cretaceous)	Chalk, limestone, and marl	Mostly microgranular calcite with minor foraminifera tests, averages 85 percent calcium carbonate, ledge forming, grayish white to white; hard lime mudstone to soft chalk in west, occasional beds with large-scale cross-stratification.	60 to 175 m	South and east of Balcones Escarpment from Del Rio to Bell County. Thickens westward. Austin Chalk includes Pflugerville, Burditt, Dessau, Jonah, Vinson, and Atco members.	Some development of caves.	Adkins, 1932; Young and Marks, 1952; Young, 1965; Pessagno, 1969.
Pflugerville Chalk Member[c] of the Austin Chalk (Upper Cretaceous)	Chalk and marl	Indurated, often phosphatic chalks with thin marl partings.	At least 10 m	South and east of Balcones Escarpment from Del Rio to north of Austin. Cut out by erosion in vicinity of New Braunfels. Included in the Austin Chalk		Young, 1963, 1965, 1977; Pessagno, 1969.
Burditt Marl Member[c] of the Austin Chalk (Upper Cretaceous)	Chalk and marl	Predominantly chalk with thin beds of marl, gray to buff colored; silty marlstones	17 to 19 m	South and est of Balcones Escarpment from Del Rio to north of Austin. Cut out by erosion in vicinty of New Braunfels. Included in the Austin Chalk.		Young and Marks, 1952; Young, 1963, 1965, 1977; Pessagno, 1969.
Dessau Chalk Member[c] of the Austin Chalk (Upper Cretaceous)	Chalk	White to light gray chalks interbedded with some marls and marlstones, becomes nodular at base.	Approximately 6 m	South and east of Balcones Escarpment from San Marcos to San Antonio. Included in the Austin Chalk.		Young, 1963, 1965, 1977; Pessagno, 1969.
Jonah Limestone Member[c] of the Austin Chalk (Upper Cretaceous)	Limestone	Fragmental limestone, packstone to grainstone, shell fragments common	11 m	East of Balcones Escarpment near Austin. Included in the Austin Chalk.		Young, 1963, 1965, 1977; Pessagno, 1969.
Vinson Chalk Member[c] of the Austin Chalk (Upper Cretaceous)	Chalk and marl	Massive, moderately resistant chalks separated by thin marly layers.	20 m	East of Balcones Escarpment near Austin. Included in the Austin Chalk.		Young, 1963, 1965, 1977; Pessagno, 1969.
Atco Chalk Member[b] of the Austin Chalk (Upper Cretaceous)	Chalk	White resistant chalk beds usually over 1 m thick with beds of blue gray and buff marls; indurated.	4 to 20m	South and east of Balcones Escarpment from San Marcos to Bell County. Included in the Austin Group.		Young, 1963, 1965, 1977; Pessagno, 1969.
Eagle Ford Group (Upper Cretaceous)	Shale, siltstone, shale and limestone	Upper part limestone and shale, limestone light yellowish brown, flaggy; lower part, siltstone and very fine-grained sandstone, light yellow to gray, laminated, flaggy, some limestone, silty, medium brown, laminated; in west lower part is shale, calcareous, dark gray.	8 to 60 m	Along Balcones Escarpment from Del Rio to Uvalde and from Austin to Belton. Thickest at Belton and in western Uvalde County and thins northeastward.		Adkins, 1932; Adkins and Lozo, 1951; Pessagno, 1969.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Boquillas Flags (Upper Cretaceous)	Shale, limestone	Consists of four units: upper unit mostly shale, silty, medium gray, interbedded with some limestone, nodular to laminar, granular, brownish gray; next unit shale, silty, medium-gray, interbedded with limestone, granular, yellowish gray; shale, silty, dark gray, interbedded with siltstone, laminated, grading upward to silty limestone; basal unit of limestone, clastic, in thin mostly cross-laminated beds that pinch and swell along strike, interbedded with siltstone, light yellowish gray to grayish orange.	50 to 70 m	In Val Verde County in Rio Grande region; gradational to Eagle Ford Group eastward. Upper two units comprise the Langtry Member, lower two units comprise the Rock Pens Member.	Some cavern development collapse areas in Val Verde County.	Adkins, 1932; Pessagno, 1969.
South Bosque Formation (Upper Cretaceous)	Shale and limestone	Mostly shale, bentonitic in lower part, soft, blocky, homogeneous, dark gray to black, weathers medium gray to yellowish brown; limestone beds scarce, more abundant downward, flaggy.	9 to 45 m	In northeasternmost part of study area; upper part of Eagle Ford Group in this location.		Adkins, 1932; Adkins and Lozo, 1951; Pessagno, 1969.
Lake Waco Formation (Upper Cretaceous)	Limestone and shale	Limestone forms top and bottom units of formation; shale, calcareous, bentonitic, interbedded with thin beds of hard limestone which are less abundant upward.	8 to 20 m	In northeasternmost part of study area; thins southward; in the northeast part of study area upper limestone beds comprise the Bouldin Flag Member, middle shale beds comprise the Cloice Shale Member, and lower limestone beds comprise the Bluebonnet Flag Member. Not deposited over most of Edwards Plateau.		Adkins and Lozo, 1951; Pessagno, 1969.
Pepper Shale (Upper Cretaceous)	Shale	Pyritic, massive, breaks with blocky fracture, thin to very thinly laminated; upon weathering develops fissility, selenite, and jarosite.	6 to 30 m	In northeasternmost part of study area, north of Travis County; transitional northward to Woodbine Formation. Not deposited over most of Edwards Plateau.		Adkins, 1932; Adkins and Lozo, 1951.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Buda Limestone (Upper Cretaceous)	Limestone	Fine-grained bioclastic, commonly glauconitic, pyritiferous, hard, massive, poorly bedded to nodular, thinner bedded to argillaceous near upper contact, light gray to pale orange; weathers dark gray to brown; burrows filled with chalky marl; in northwest, aphanitic, compact, nodular, mollusc-fragment bearing, with marly interbeds, gray, yellowish gray, basal beds typically marly, nodular limestone and thin yellow marl.	0 to 30 m	Throughout area; caps ridges in many localities; in northwest, finely mottled pattern on aerial photographs, grassy "racetrack" forms periphery. Thickens eastward from Val Verde County to San Antonio.	Some karstification.	Hixon, 1959; Martin, 1967; Rose, 1968, 1972.
Del Rio Clay (Upper Cretaceous)	Clay	Calcareous and gypsiferous becoming less calcareous and more gypsiferous upward, pyrite common, blocky, medium-gray, weathers light gray to yellowish gray; some thin lenticular beds of highly calcareous siltstone; in northeast, some thin limestone beds composed mostly of pelecypods, some beds of ocherous hematite.	12 to 60 m	Generally along south and east margins of study area, patchy; thickens west of San Antonio, thins onto San Marcos Platform, feathers out in southwest and northwest. Referred to as "Grayson Marl" in northeast. Not deposited over most of the Edwards Plateau where Buda Limestone rests on Segovia Formation.		Martin, 1961; Rose, 1968, 1972.
Georgetown Formation (Lower and Upper Cretaceous)	Limestone and marl	Mostly limestone, fine grained, argillaceous, nodular, moderately indurated, light gray; some limestone, hard, brittle, thick bedded, white; some shale, calcareous, soft, light gray to yellowish gray.	3 t0 25 m	Along eastern margin of study area in Balcones Fault Zone. Thins southward. Includes Main Street, Paw Paw, Weno, Denton, Fort Worth, and Duck Creek Members in northeast.	Locally cavernous.	Wilbert, 1963, 1967.
Main Street Member of the Georgetown Formation (Lower Cretaceous)	Limestone	Medium grained, chalky, some 2-3 m interbeds of calcareous shale, thin bedded to massive, distinctly bedded to wavy bedded and nodular; weathers light gray to white.	7 to 11 m	In northeast along Balcones Fault Zone. South to Comal County. Included in Georgetown Formation. Not continuously deposited over Edwards Plateau.		Wilbert, 1963, 1967.
Paw Paw Member[c] of the Georgetown Formation (Lower Cretaceous)	Marl	Calcareous, soft ledge-forming limestone bed near middle, unit as a whole recessive.	Up to 3 m	In northeast along Balcones Fault Zone, thins southward. Absent south of Waco. Included in Georgetown Formation.		Wilbert, 1963, 1967.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Weno Member[c] of the Georgetown Formation (Lower Cretaceous)	Limestone	Some very thin marl interbeds, thin- to medium-bedded, white to grayish yellow, basal 0.6-1.2-m resistant ledge forms uplands.	2 to 14 m	Northeast of Round Rock along Balcones Fault Zone. Included in Georgetown Formation.		Wilbert, 1963, 1967.
Denton Member[c] of the Georgetown Formation (Lower Cretaceous)	Clay	Calcareous, argillaceous limestone in upper part, brownish grayish yellow.	1 to 3 m	In Hays County and northeast along Balcones Fault Zone. Included in Georgetown Formation.		Wilbert, 1963, 1967.
Fort Worth Member[c] of the Georgetown Formation (Lower Cretaceous)	Limestone and marl	Limestone, chalky, medium-grained, fairly hard, nodular, bluish white to bluish gray, interbedded with light gray marl.	7 to 11 m	In Hays County and northeast along Balcones Fault Zone. Included in Georgetown Formation.		Wilbert, 1963, 1967.
Duck Creek Member[c] of the Georgetown Formation (Lower Cretaceous)	Limestone and marl	Limestone, medium-bedded, nodular to wavy bedded, gray, interbedded with marly clay; harder and more resistant than Fort Worth Member.	7 to 9 m	In northeast along Balcones Fault Zone. Absent south of Austin. Included in Georgetown Formation.		Wilbert, 1963, 1967.
Segovia Formation (Lower Cretaceous)	Limestone and dolomite	In upper part, medium- to coarse-grained, cherty, resistant, light-gray, miliolid, shell fragment, rudistid limestone; in middle part, dolomite, medium brownish gray, porous, massive to thin-bedded, cherty, collapse breccia (Allen Ranch Breccia); in lower part light yellowish-gray miliolid limestone and marl and marly limestone; in north, uppermost beds include one or more thick limestone units, each forming prominent ledge or "caprock"; key beds include a rounded resistant ledge of biosparite (Black Bed), a distinctively open slope with petrified wood and chert nodules (Orr Ranch Bed), a caprinid zone, and one or more resistant beds of whole-oyster biomicrite with abundant silicified Gryphaea (Gryphaea Bed).	50 to 115 m	Extensive outcrops in uplands north of Balcones Fault Zone and west of Llano Basin; thickens southward. Includes Allen Ranch Breccia and Burt Ranch Member. Included in Edwards Group. Upper part is equivalent to Main Street Member of Georgetown Formation.	Karsted and in part cavernous; forms caprock over caves. Together with Fort Terrett Formation and Trinity Group forms unconfined Edwards-Trinity aquifer.	Moore, 1967; Rose, 1968, 1969, 1972.
Allen Ranch Breccia[c] (Lower Cretaceous)	Dolomite, limestone, and collapse breccia.	Brownish gray, thin- to thick-bedded medium crystalline, homogenous dolomite interbedded with light gray, brittle, siliceous micrite.	6 to 7 m	Local zones of collapse. Included in Segovia Formation.	Solution-Collapse structures.	Rose, 1968, 1972.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
West Nueces Formation (Lower Cretaceous)	Limestone	Upper part limestone, fine-grained, massive, miliolid and mollusc-bearing; lower part limestone, nodular, oysters, and other molluscs common. Basal Transgressive Unit correlative with Basal Nodular Member of Fort Terrett Formation to north and Bull Creek Limestone Member of Walnut Formation to northeast.	43 m + or -	Northeastern Kinney County and western Uvalde County. Included in Edwards Group.	Karsted.	Lozo and Smith, 1964.
Person Formation[c] (Lower Cretaceous)	Limestone, dolomite, and dolomitic Limestone	Consists of Cyclic Member, Marine Member, Leached Member, Collapsed Member, and Regional Dense Member (see below). Also included in Edwards Group (see below).	60 to 70m	Crops out in Balcones Fault Zone in eastern part of San Marcos Platform. Correlative with Segovia Formation to west. Included in Edwards Group. Mapped as Edwards Limestone.	Cavernous, karsted; recharge zone of Edwards artesian aquifer; host rock to aquifer in subsurface.	Rose, 1968, 1972.
Cyclic Member[c] of the Person Formation (Lower Cretaceous)	Limestone and dolomitic limestone	Medium to dark gray and brown, burrowed, wispy intramicrite and mollusc biomicrite, in part dolomitic; micrite matrix altered partly to microspar; intraclasts are broken stromatolitic crusts and soft mud lumps, many shells have been leached and later filled with sparry calcite; few thin collapsed zones; black chert nodules near base of bed; upper boundary is abrupt, irregular, with truncated shells and fractures.	18 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Uppermost member of Person Formation.	Caves and dolines filled with Georgetown-equivalent rock and angular Edwards-type lithoclasts.	Rose, 1968, 1972.
Marine Member[c] of the Person Formation (Lower Cretaceous)	Limestone	Medium brown, wispy, medium- to coarse-grained mollusc biosparite and biomicrite; micrite matrix partly altered to microspar; skeletal components occur whole or as sand-sized fragments; much shell material has been dissolved and voids later filled with sparry calcite; scattered small patches of blue-gray chalcedony replace shells; chert nodules in upper part; distinct current streaks and cross-bedding; becomes darker, finer grained in bottom 1 m.	17 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Included in Person Formation.	Cavernous, karsted.	Rose, 1968, 1972.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Leached Member[c] of the Person Formation (Lower Cretaceous)	Dolomite	Light to medium brownish gray, massive, burrowed, wispy, fine-grained, fine- to medium-crystalline, dolomitized pelleted mollusc-fragment micrite and biomicrite; sparry calcite pseudomorphs after gypsum crystals and nodules, also some spar-filled collapse cracks.	12 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Included in Person Formation.	Locally cavernous, karsted.	Rose, 1968, 1972.
Collapsed Member[c] of the Person Formation (Lower Cretaceous)	Dolomite and dolomitic limestone	Dark brown and medium gray, very coarse-grained collapse breccia and pelleted mollusc biomicrite; lithoclasts composed of broken stromatolitic crusts, laminated micrite, leached dolomite, soft mud lumps; upper part with spar-filled collapse cracks and calcite pseudomorphs after gypsum.	13 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Correlative with Allen Ranch Breccia to west. Included in Person Formation.	Solution-Collapse structures	Rose, 1968, 1972.
Regional Dense Member[c] of the Person Formation (Lower Cretaceous)	Limestone	Argillaceous, dark brownish gray, hard, compact, wispy, pyritic argillaceous micrite; scattered pellets and small oysters; in lower 2 to 3 m miliolids and composite clasts become more abundant downward; in basal 60 cm rock is rotten, granular, very wispy, carbonaceous and pyritic.	4 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Correlative with part of Kiamichi shale to northeast and part of Burt Ranch Member of the Segovia Formation to the west.	Forms caprock for some caves in Glen Rose Formation. Included in Person Formation.	Rose, 1968, 1972.
Kainer Formation[c] (Lower Cretaceous)	Limestone and dolomitic limestone	Consists of Grainstone Member (see below), Kirschberg Evaporite (see above), Dolomitic Member (see below), and Basal Nodular Member (see above). Also included in Edwards Group (see below).	120 m + or -	Crops out in Balcones Fault Zone in eastern part of San Marcos Platform. Generally correlative with Fort Terrett Formation in west. Included in Edwards Group. Mapped as Edwards Limestone.	Cavernous, karsted; Recharge zone of Edwards artesian aquifer; host rock to aquifer in subsurface	Rose, 1968, 1972.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Grainstone Limestone Light to medium brown, generally medium-grained, 43 m Balcones Fault Zone in eastern Locally Rose, 1968, Member[c] of the Kainer Formation (Lower Cretaceous)	Limestone	Light to medium brown, generally medium-grained, well sorted, current laminated, miliolid intrasparite; many clasts are coated aggregates; some micrite matrix, many clasts and miliolids altered to microspar; few streaks of larger mollusc shells; scattered dolomitic streaks in lower 9 to 12 m; few calcite pseudomorphs after anhydrite nodules.	43 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Correlative with Burt Ranch Member of Segovia to west.	Locally cavernous, karsted.	Rose, 1968, 1972.
Dolomitic Member[c] of the Kainer Formation (Lower Cretaceous)	Dolomitic limestone	Medium to dark brownish gray, burrowed, fine- to medium-grained, moderately sorted, miliolid biosparite and intrasparite; clasts are soft pellets, stromatolitic hard clasts, and coated aggregates; wispy burrowed micrite, anhydrite nodules, anhydrite filled cracks, layered celestite.	52 m + or -	Balcones Fault Zone in eastern part of San Marcos Platform. Correlative with Dolomitic and Burrowed Members of Fort Terrett Formation to west.	Locally cavernous, karsted.	Rose, 1968, 1972.
Kiamichi Shale[c] (Lower Cretaceous)	Clay, shale, and limestone	Clay and shale, calcareous, silty, yellowish brown; limestone, marly, thin nodular to wavy beds.	Up to 5 m	Narrow outcrop in extreme northeast corner of study area, north of Georgetown. Correlative with Regional Dense Member of Person Formation to southwest. Included in Edwards Group.		Adkins, 1932; Shelburne, 1959; Young, 1967; Bishop, 1967.
Edwards Formation (Lower Cretaceous)	Limestone	Aphanitic to coarse-grained, abundant chert, medium gray to grayish brown, fossils include rudists, miliolids, and shell fragments; solution zones and collapse breccia common; chert nodules and plates; massive to thin-bedded, hard, brittle; honeycombed; locally bored; locally dolomitic; resistant, generally forms scarp, at least in basal part.	5 to 150 m	Along Balcones Fault Zone from northeast Medina County to San Marcos and from north of Austin to Bell and Lampasas Counties. Thins northward. Includes Person and Kainer Formations south of Austin and equivalent to Kainer Formation north of Austin. Included in Edwards Group.	Cavernous and karsted. Recharge zone and host formation for Edwards artesian aquifer.	Lozo, 1959; Young, 1967; Rose, 1968, 1972.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Comanche Peak Formation (Lower Cretaceous)	Limestone	Fine to very fine-grained, fairly hard, nodular, numerous shale partings, light gray, weathers white, extensively burrowed, burrow fillings slightly coarser and darker, typically crops out in scarp face beneath Edwards Formation; argillaceous in northwest.	4 to 30 m	Patchy outcrops in northeast of study area. Feathers out southward near Williamson-Travis County line.		Moore, 1961a, b, 1964; Young, 1967; Hayward and Brown, 1967.
Walnut Formation (Lower Cretaceous)	Clay, claystone, limestone, marl, and shale	Clay, calcareous; limestone, chalky, marly, nodular, locally hard, thick-bedded, a few hard beds with sparry calcite, massive beds common in lower part; shale as thin beds common in upper part, sandy, light to dark gray, grayish yellow; marl, sandy in part, thin-bedded, yellow; claystone, calcareous, locally sandy, in part nodular, light brown to gray.	6 to 53 m	In northeast part of study area. Feathers out to south. Includes Upper Marl Member, Keys Valley Marl Member, Whitestone Limestone Member, Cedar Park Limestone Member, Bee Cave Marl Member, and Bull Creek Limestone Member.	Intersected by some cavern development.	Lozo, 1959; Moore, 1959, 1961a, 1964; Moore and Martin, 1966; Young, 1967.
Upper Marl Member[c] of the Walnut Formation (Lower Cretaceous)	Limestone and marl	Nodular, fossiliferous, and clayey biomicrite, flaggy, miliolid, mollusc and oyster fragments; lower part contains echinoids.	2 to 16 m	Narrow outcrops in northeast part of study area in Bell County. Included in Walnut Formation.		Moore, 1964; Young, 1967.
Keys Valley Marl Member of the Walnut Formation (Lower Cretaceous)	Marl	Soft, white; marine fossils include pelecypods, ammonites, gastropods, and echinoids, clayey nodular biomicrite and micrite.	0 to 16 m	Crops out in northwest Williamson County. Correlative in part with Whitestone Limestone Member. Included in Walnut Formation.		Moore, 1964; Young, 1967.
Whitestone Limestone Member[c] of the Walnut Formation (Lower Cretaceous)	Limestone, grainstone, mudstone	Upper part, fossiliferous oosparite, mollusc shell fragments; rudistid shell-fragment biomicrite; fossiliferous pellet intrasparite at base.	12 to 20 m	Narrow outcrops in northeast part of study area in Burnet, Williamson, and Travis Counties. Included in Walnut Formation.		Moore, 1964; Moore and Martin, 1966.
Cedar Park Limestone Member of the Walnut Formation (Lower Cretaceous)	Limestone	Fossiliferous nodular micrite; scattered mollusc fragments; shell-fragment biomicrite, algae, echinoid spines and miliolids; clayey biomicrite, intrasparite flags.	7 to 15 m	Crops out in western Williamson and Travis Counties and eastern Burnet County. Included in Walnut Formation.		Moore, 1964; Young, 1967.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Bee Cave Marl Member of the Walnut Formation (Lower Cretaceous)	Marl	Clayey fossiliferous micrite and fossiliferous marl south of Williamson County; brown fossiliferous clay and micrite flags in northern Williamson County, Bell, Lampasas, and Coryell Counties; intraclastic clayey biomicrite in southern Williamson County.	0 to 12 m	Crops out in northwestern Williamson County and eastern Burnet County and along the eastern margin of study area as far south as Comal County. Thins to west, south, and southwest. Included in Walnut Formation.	Intersected by some cavern development.	Cuyler, 1939; Moore, 1964; Young, 1967.
Bull Creek Limestone Member[c] of the Walnut Formation (Lower Cretaceous)	Limestone	Thin- to medium-bedded, hard, well sorted biomicrite, intramicrite, and fossiliferous intrasparite.	0 to 12 m	Crops out in northwestern Williamson County and eastern Burnet County and along eastern margin of study area as far south as Comal County. Thins to north, west, and south from maximum thickness in Williamson County. Included in Walnut Formation.	Intersected by some cavern development.	Moore, 1964; Young, 1967.
Paluxy Sandstone (Lower Cretaceous)	Sandstone and claystone	ery fine- to medium-grained, unindurated quartz sand; finely laminated or cross-bedded, nonfossiliferous; friable, light gray to red; in part silty, argillaceous, calcite cemented and hard, some thin interbeds of gray shale and limestone, pyrite nodules and concretions.	0 to 20 m	Narrow outcrop bands in northeast part of study area in Bell and Travis Counties. Thins southward. Sandstone outcrops in Burnet and Lampasas Counties are the Oatmeal Sandstone.		Moore, 1964;
Glen Rose Formation (Lower Cretaceous)	Limestone, dolomite, marl, and Clay	Alternating resistant and recessive beds forming stairstep topography; limestone, aphanitic to fine-grained, hard to soft and marly, white, light gray to yellowish gray, arenaceous in north; dolomite, fine-grained, porous, yellowish brown; marine megafossils include molluscan steinkerns, rudistids, oysters, and echinoids; subdivided into two units by Corbula bed.	60 to 275 m	Extensive outcrops in dissected margins of southern and eastern Edwards Plateau. Thins northwestward.	Cavernous and slightly karsted. With Edwards Group forms extensive Edwards-Trinity unconfined aquifer.	Young, 1967;. Stricklin and others, 1971.
Upper Member of Glen Rose Formation (Lower Cretaceous)	Limestone, dolomite, and marl	Relatively thinner bedded, more dolomitic, and less fossiliferous.	120 m + or -	In dissected margins of southern and eastern Edwards Plateau.	Cavernous and karsted.	Young, 1967;. Stricklin and others, 1971.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Lower Member of Glen Rose Formation (Lower Cretaceous)	Limestone, dolomite, and marl	More massive, contains rudistid reefs and thick Corbula steinkern interval at top.	150 m + or -	In dissected margins of southern and eastern Edwards Plateau.	Cavernous and karsted.	Young, 1967; Stricklin and others, 1971.
Hensell ´ Sand (Lower Cretaceous)	Sandstone and limestone	Mostly fine-grained, friable to well cemented, argillaceous, calcareous, light gray to brownish gray; upper half dolomitized limestone, sandy, glauconitic, honeycombed, nodular; lower half mostly sandstone, fine-grained, argillaceous, calcareous, contains white, hard, siliceous geodes up to 20 cm in diameter; some hard limestone contains large oysters.	14 m + or -	Narrow outcrops along bottoms of major stream valleys in Burnet, Williamson, Blanco, Hays, Comal, and Kendall Counties	Some caverns developed perched on Hensell Sand.	Lozo and Stricklin, 1956; Fisher and Rodda, 1966, 1967a; Boone, 1968; Stricklin and others, 1971.
Travis Peak Formation (Lower Cretaceous)	Conglomerate, sandstone, and limestone	Conglomerate, sandy, composed of pebbles, cobbles, and boulders of Paleozoic limestone, dolomite, chert, and sandstone, calcite cemented, red to¨variable white; sandstone, mostly quartz, conglomeratic, silty, locally composed of fossil fragments, red to white; some limestone, silty, sandy, white.	Up to 54 m, variable	Narrow outcrops in Lampasas, Burnet, Williamson, Travis, and Blanco Counties, and in deeply incised stream valleys in Hays, Comal, and Kendall Counties. Includes Cow Creek Limestone, Hammett Shale and Sycamore Sand Members. Travis Peak Formation no longer used as stratigraphic unit.		Lozo and Stricklin, 1956; Fisher and Rodda, 1966, 1967a; Young, 1967; Boone, 1968; Stricklin and others, 1971.
Cow Creek Limestone Member of the Travis Peak Formation (Lower Cretaceous)	Limestone	Massive, off-white, in part honeycombed, fossiliferous; upper part, ouquine with inclined and cross-bedding; lower part, fine silty quartz calcarenite and massive with oyster beds.	23 m + or -	Narrow outcrops in Lampasas, Burnet, and Williamson Counties and along Pedernales River in Blanco and Travis Counties, along Blanco River in Hays County, along Guadalupe River in Kendall and Comal Counties. Thickens eastward Included in Travis Peak Formation.		Lozo and Stricklin, 1956; Fisher and Rodda, 1966; Young, 1967; Stricklin and others, 1971; Stricklin and Smith, 1973.
Hammett Shale Member of the Travis Peak Formation (Lower Cretaceous)	Shale	Light brown to gray claystone, dolomitic limestone near top; stringers of quartz sandstone and siltstone near base; uppermost part contains fossiliferous dolomitic limestones.	0 to 18 m	Crops out in northeast part of study area in Travis and Burnet Counties. Included in Travis Peak Formation.		Lozo and Stricklin, 1956; Stricklin and others, 1971.

Table 13: continued

Formation or Member[b]	Type of Rock	Lithology	Thickness	Extent of Outcrop	Significance to this Study	Primary References
Sycamore Sand Member[c] of the Travis Peak Formation (Lower Cretaceous)	Sandstone conglomerate, and claystone	Irregularly bedded, glauconitic, red; quartz sand most abundant gradation from coarse to fine in pward direction, lenticular channel deposits, pebbles and boulders derived from Paleozoic and older rocks of the Llano Basin, also pebbles and cobbles of dolomite, limestone, and chert.	0 to 46 m	Crops out in Colorado and Pedernales River valleys, west and northwest of Austin.		Fisher and Rodda, 1966; Young, 1967; Boone, 1968; Stricklin and others, 1971.
Pearsall Formation[c] (Lower Cretaceous)	Shale	Light to dark gray, silty, calcareous.	30 to 40 m	In central Texas Pearsall correlates with units of the Cow Creek Limestone and Hammett Shale Members of the Travis Peak Formation. Name generally applied to subsurface part.		Hayward and Brown, 1967; Boone, 1968.
Sligo Limestone[c] (Lower Cretaceous)	Limestone	Upper unit of oolitic lime packstone and grainstone, dolomitic, burrowed; lower unit of laminated dolomite, siltstone, and gray lime mudstone and wackestone.	20 to 70 m	Entirely subsurface beneath study area. Correlates in part with Sycamore Sand Member of the Travis Peak Formation.		Hayward and Brown, 1967; Stricklin and others, 1971.
Hosston Sand[c] (Lower Cretaceous)	Sandstone	Fine- to medium-grained immature to submature, friable, compact, orthoquartzitic, interbedded with sandy mud and muddy, pebbly, sandstone; sands, thin- to massively bedded, cross-bedding common in conglomeratic parts.	10 to 30 m	Entirely subsurface beneath study area. Correlative with Sycamore Sand Member of Travis Peak Formation.		Fisher and Rodda, 1966; Hayward and Brown, 1967; Boone, 1968.
Antlers Sand (Lower Cretaceous)	Sandstone, conglomerate, siltstone, claystone, and quartzite	Sandstone, fine- to very fine-, locally coarse-grained, argillaceous, massive, locally cross-bedded and ripple marked, gray, yellowish brown, red; conglomerate, sandy, caliche cemented, pebbles of chert and quartz, large boulders of limestone near base; siltstone, thin beds, gray, pink; claystone, sandy, calcareous, blocky, brown to purplish pink, green to gray.	30 to 60 m, variable	Crops out along northern boundary of area below Fort Terrett Formation. Correlative and gradational with Glen Rose and Hensell Formations and members of the Travis Peak Formation in the subsurface.		Fisher and Rodda, 1966; Hayward and Brown, 1967; Boone, 1968.

APPENDIX C
CAVES OF THE EDWARDS PLATEAU

CATALOGING CAVES OF THE EDWARDS PLATEAU

Records of known caves have been kept for all states by the National Speleological Society since its founding in 1939 and by its many chapters and regional organizations. The number of listed caves for Texas in the early 1940's was meager (Morgan, 1943, 1944); however, interest increased following the publication of a special issue of the National Speleological Society Bulletin on the caves of Texas later in that decade (Craun, 1948; White, 1948a,b; anonymous, 1963a). A record of progress in cataloging the caves of Texas and those of the Edwards Plateau in particular is given in Table 14 and Figure 172. Although the intensity of exploration and surveying of caves in Texas varies from year to year, the trend of growth portrayed in these records is expected to continue for some time.

Table 14: Record of reported caves of the Edwards Plateau by county[a]

County	1944	1948	1958	1961	1962	1966	1979
Bandera				9	20	22	31
Bell		6	6	10	14	29	29
Bexar	1	4	4	27	64	66	77
Blanco		2	2	4	6	6	13
Burnet	1	6	12	15	24	33	45
Comal		6	6	27	36	85	95
Concho							
Coryell						16	17
Crockett		1	1	15	17	17	19
Edwards		4	18	37	42	50	81
Falls					1	1	1
Gillespie		1	1	17	17	18	18
Guadalupe							
Hays	2	11	11	26	34	36	77
Irion			3	3	4	4	4
Kendell	2	14	28	62	65	65	72
Kerr		7	7	34	39	47	58
Kimble		1	1	9	14	15	47
Kinney		1	1	4	12	15	22
Lampasas				1	15	45	40
Llano				4	4	4	17
Mason		2	2	18	19	26	26
McCulloch				4	12	13	13
McLennen						1	2
Medina		3	3	27	29	30	38
Menard				2	2	9	9
Mills					1	1	1
Reagan							
Real		2	2	13	14	17	32
San Saba	1	7	7	20	67	121	197
Schleicher				1	6	6	8
Sutton				10	33	37	46
Terrell		6	6	6	6	10	13
Tom Green							3
Travis	1	7	13	41	85	117	149
Uvalde		9	9	26	62	63	64
Val Verde		30	42	49	58	100	127
Williamson		10	15	39	47	77	81
				Totals			
Edwards Plateau	8	140	200	560	869	1202	1572
Remainder of Texas	1	45	50	85	162	186	497
All of Texas	9	185	250	645	1031	1388	2069
Source[b]	(1)	(2)	(3)	(4)	(5)	(6)	(7)

[a]See also Figure 172.

[b]Sources: (1) Morgan (1943, 1944)
(2) National Speleological Society Bulletin, v. 10, 1948
(3) Widener (1958)
(4) Reddell (1961a)
(5) anonymous (1962a)
(6) Reddell and Smith (1966)
(7) J. Reddell, written communication

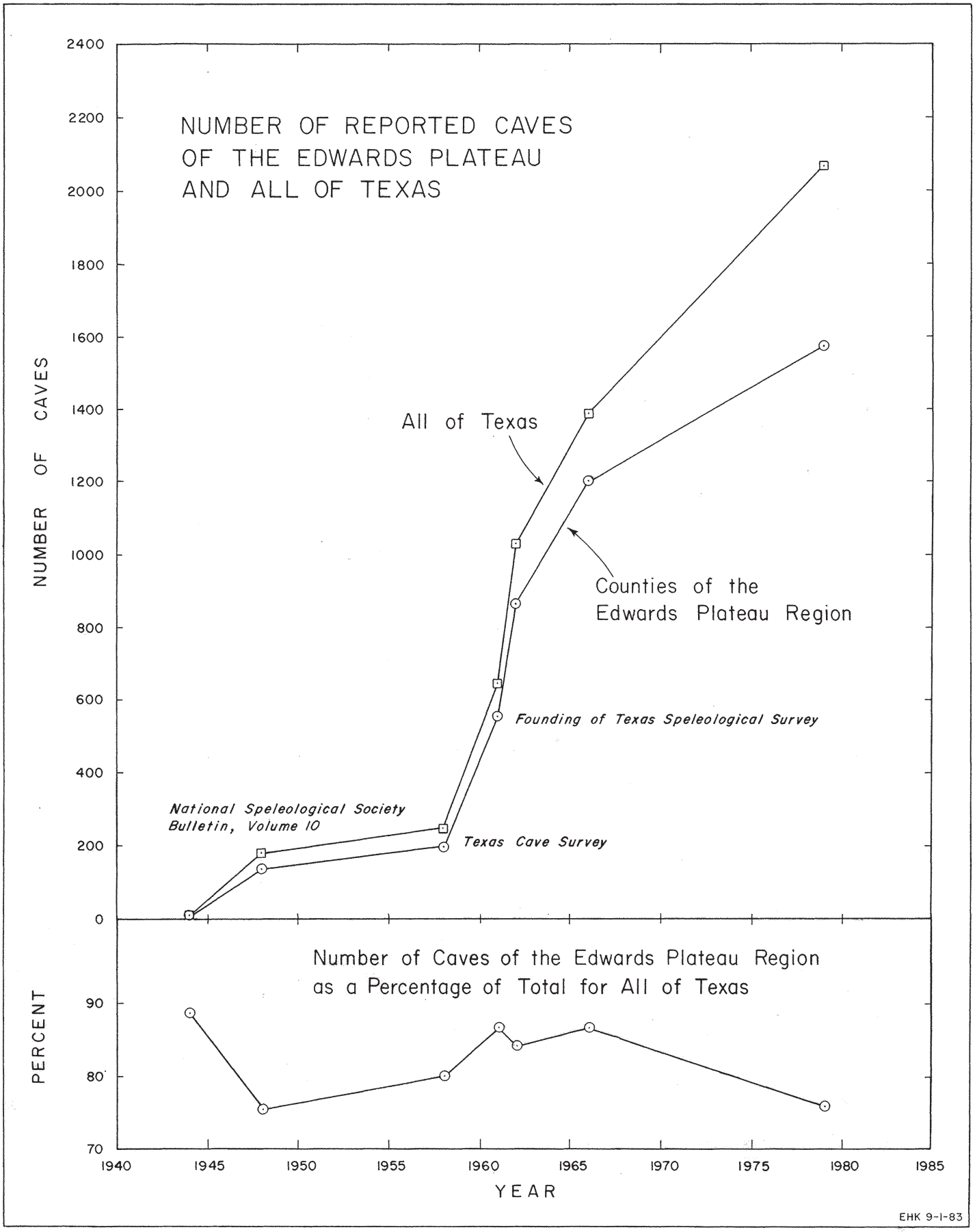


Figure 172: Number of reported caves of the Edwards Plateau and all of Texas, 1944 to 1979. Lower graph indicates percentage of total caves of Texas that are within counties listed in Table 1. Number of caves of individual counties are given in Table 14.

SIGNIFICANT CAVES OF THE EDWARDS PLATEAU

The majority of caves in the Edwards Plateau are small (less than 25 m in length and less than 25 m in depth). Those that exceed either of these dimensions and for which sufficient documentation is available are shown on Plate 1. Caves that exceed 300 m in surveyed length are numbered on Plate 1 and listed in Table 15.

Table 15: Significant Caves of the Edwards Plateau[a]

County	Number[b]	Cave[c]	Area[d]
Bandera	1	Station C Cave	Mountain Home-Leakey
Bexar	1	Robber Baron's Cave	San Antonio
	2	Wurzbach's Bat Cave	San Antonio
	3	Fair Hole	Boerne
Blanco	1	Buffalo Cave	Johnson City
Burnel	1	*Longhorn Cavern	Burnet
	2	Bad Air Cave	Smithwick
	3	Beaver Creek Caverns	Bend
Comal	1	*Natural Bridge Caverns	New Braunfels
	2	*Bracken Bat Cave	New Braunfels
	3	Dinosaur Cave	New Braunfels
	4	Honey Creek Cave	Boerne
	5	Bartels Cave	Boerne
Crockett	1	0-9 Well	Suggs
	2	Water Cave	Big Lake
Edwards	1	Dunbar Cave	Carta Valley
	2	Deep Cave	Carta Valley
	3	Dragoo Cave	Carta Valley
	4	Pyramid Cave	Carta Valley
	5	Midnight Cave	Carta Valley
	6	Schulze Cave	Telegraph
Hays	1	McCarty Cave	Austin-San Marcos
Irion	-	**Arden Cave	Middle Concho River
Kendall	1	Cave Without a Name	Boerne
	2	Spring Creek Cave	Boerne
	3	Prassell Ranch Cave	Boerne
	4	Alzafar Cave	Boerne
	5	Cascade Caverns	Boerne
	6	Cascade Caverns Sinkhole	Boerne
Kerr	1	Stowers Cave	Mountain Home-Leakey
Kimble	1	Fleming Bat Cave	Telegraph
	2	Paint Creek cave	Telegraph
Kinney	1	Palace Cave	
	2	Green Cave	Brackettville
	3	Kickapoo Cave	Brackettville
	4	Webb Cave	Brackettville
Lampasas	1	Battery Cave	Bend
Llano	1	*Enchanted Rock Cave	Enchanted Rock
Medina	1	Valdina Farms Sinkhole	Sabinal
	2	Marguerite Cave	Sabinal
	3	Finger Cave	Sabinal
Menard	1	*Powell-Neel Cave System	Menard
	2	*Silver Mine Sink	Menard
Real	1	Red Arrow Cave	Vance
	2	Bradford Cave	Vance
	3	Cave of the Lakes	Mountain Home-Leakey
	4	Dan Auld Ranch Cave No. 1	Mountain Home-Leakey
	5	Dan Auld Ranch Cave No. 2	Mountain Home-Leakey
	6	Perry Water Cave	Mountain Home-Leakey
	7	Turkey Pens Cave	Vance
San Saba	1	*Gorman Cave	Bend
	2	*Clark's Branch Well Cave	Bend
	3	Four Bat Cave	San Saba
Schleicher	1	Cave Y	Eldorado
Sutton	1	*Caverns of Sonora	Sonora
	2	*Felton Cave	Sonora
	3	*Silky Cave	Sonora
	4	Alma's Cave	Sonora
	5	Word Cave	Sonora
	6	Harrison Cave	Sonora
Terrell	-	**Sorcerer's Cave	Dryden
Travis	1	*Airman's Cave	Austin-San Marcos
	2	*Cave X	Austin-San Marcos
	3	Austin Caverns	Austin-San Marcos
Uvalde	1	Indian Creek Cave	Uvalde
	2	Rambie's Cave	Uvalde
	3	Frio Bat Cave	Sabinal
Val Verde	1	*Langtry Lead Cave	Langtry
	2	*Emerald Sink	Langtry
	3	*Langtry Quarry Cave	Langtry
	4	Fern Cave	Devils River
	5	Fawcett Cave	Carta Valley
	6	H. T. Meirs Cave	Carta Valley
	7	Diablo Cave	Del Rio
	8	Four-Mile Cave	Del Rio
	9	Oriente Milestone Molasses Bat Cave	Del Rio
	10	Dondole Cave	Pandale
	11	Oberkampf Ranch Cave	Pandale
	12	Marshall Bat Cave	Pandale
Williamson	1	*Inner Space Cavern	Georgetown
	2	Steam Cave	Georgetown
	3	Beck Ranch Cave	Georgetown
	4	Bat-Well	Georgetown
	5	Cobb Caverns	Georgetown

[a]A list of well documented caves exceeding 300 m in surveyed length. From data compiled by J. R. Reddell and A. R. Smith, 1979-1981, and from reports of the Texas Speleological Survey (Table 4).

[b]Number within county refers to location on Plate 1

[c]An asterisk indicates that this cave was investigated in this study. A double asterisk indicates that the cave lies outside of the area of study.

[d]Names of caves correspond to those shown in Plate 1 or Figure 171 and listed in Table 8 or Table 12 respectively.

LONG AND DEEP CAVES OF TEXAS

The longest and deepest caves of Texas are listed in Tables 16 and 17 respectively. All but 2 of the 17 longest and 6 of the 19 deepest are within the area of study. Lengthy and deep caves are products of favorable, speleogenetic conditions that existed at specific localities. For those caves investigated in this study, the particular factors responsible for their development are discussed in Part II.

Table 16: Long Caves of Texas[a]

Rank	Cave	County	Length[b] (m)	Extent[c] (m)	Reference
1	Powell-Neel Cave System	Menard	19,500+	3050+	Texas Speleological Survey (manuscript maps)
2	Honey Creek Cave	Comal	3250	4578	Poole (1983)
3	Indian Creek Cave	Uvalde	6710	1284	Reddell (1964b), Pate (1979)
4	Inner Space Cavern	Williamson	4600+	644	Pate (1979)
5	Cave Without a Name	Kendall	4310	1811	Kastning (1978), Pate (1979)
6	Airman's Cave	Travis	3410	980	Russell (1975b), Pate (1979)
7	Longhorn Caverns	Burnet	2350	1100	Pate (1979)
8	Sorcerer's (Adams) Cave[d]	Terrell	2690	1369	Veni (1980a), Poole (1981)
9	Natural Bridge Caverns (North Caverns only)	Comal	2625	923	Pate (1979)
10	Prassell Ranch Cave	Kendall	2618	1880	R. T. Bartholomew (personal communication) Pate (1979)
11	River Styx Cave[d]	King	2557	883	Kastning (1976, 1977d) Pate (1979)
12	Stowers Cave	Kerr	2390	320	R. T. Bartholomew (personal communication) Pate (1978)
13	Caverns of Sonora	Sutton	2200	656	Kastning (1983a)
14	Diablo Cave	Val Verde	2060	?	D. Pate (personal communication)
15	Felton Cave	Sutton	2050	350	Kastning (1983a)
16	Harrison Cave	Sutton	1830	?	D. Pate (personal communication)
16	Robber Baron Cave	Bexar	1341	104	Waters (1979, 1980)

[a]This table is based on information, published and otherwise, available in 1983. Tabulated values change often as exploration continues and as caves are more accurately surveyed.
[b]Length represents total, mapped length of all passages as projected onto a horizontal plane.
[c]Extent represents horizontal distance in a straight line between those two points in the mapped cave that are furthest apart.
[d]Cave is not in the area of study.

Table 17: Deep Caves of Texas[a]

Rank	Cave	County	Depth[b] (m)	Reference
1	Sorcerer's (Adams) Cave[c]	Terrell	170.0	Veni (1980c)
2	Wizard Well[c]	Terrell	118.3	Melot (1982)
3	Langtry Lead Cave	Val Verde	113.1	Pate (1979)
4	Plateau Cave[c]	Culberson	103.6	Pate (1979)
5	0-9 Water Well	Crockett	101.1	Veni (1980b)
6	Blowhole	Edwards	100.9	Pate (1979)
7	H. T. Meirs	Val Verde	100.0	Pate (1979)
8	Helm's West Well[c]	El Paso	96.0	Pate (1979)
9	400 Foot Cave[c]	Brewster	95.1	Pate (1979)
10	Devil's Sinkhole	Edwards	93.5	Pate (1979)
11	Emerald Sink	Val Verde	91.4	Pate (1979)
12	Deep Cave	Edwards	91.1	Pate (1976, 1977)
13	Langtry Quarry Cave	Val Verde	82.9	Pate (1976, 1977)
14	Fisher's Fissure	Val Verde	81.0	Pate (1976, 1977)
15	Natural Bridge Caverns	Comal	76.2	Pate (1977)
16	Abominable Sinkhole	Val Verde	71.0	Pate (1976)
17	Montgomery Gypsum Cave[c]	Terrell	68.6	Pate (1976)
18	Harrison Cave	Sutton	63.1	Pate (1976)
19	Whiteface Cave	San Saba	61.0	Pate (1976)

[a]This table is based on information, published and otherwise, available in 1983. Tabulated values change often as exploration continues and as caves are more accurately surveyed.
[b]Depth represents the difference in elevations fo the highest and lowest points in the cave.
[c]Cave is not in the area of study.

APPENDIX D
TECHNIQUES OF MAPPING CAVES

Techniques of mapping caves are fairly standardized throughout the world and have been described in detail in various manuals and papers (e.g. Davies, 1947; Swinnerton, 1950; Butcher, 1950, 1962; Edwards, 1958; Brod, 1962; Butcher and Railton, 1966; Bassham, 1971; Hosley, 1971; McClurg, 1973, p. 162-169; Freeman, 1975; Ellis, 1976a,b; Nieland, 1975, 1978; Daunt-Mergens, 1981; Chabert and Watson, 1981). Speleologists working in Texas have developed several innovative modifications to the techniques (Lindsley, 1964; Bartholomew, 1982; Jasek, 1982; Knox, 1982; Schroeder, 1982; Warton, 1982). An extensive bibliography on surveying caves has been compiled by Ellis (1976a).

GENERAL PROCEDURES

Surveying

A survey team of two to five participants progresses from point to point (station to station) through the cave, along passages and within rooms. Distance, azimuth (or bearing), and inclination along a line of sight from one station to the next are measured and recorded. A notetaker sketches plan, profile, and cross-sectional views of the passages and rooms and records dimensions of walls, heights of ceilings, elevations of floors, and internal details such as blocks of breakdown, streams, sediments, pits, ledges, domes, speleothems, joints, faults, and other phenomena of interest. Surveys of side-passages are connected to those of the main passages, thereby forming a single network of traverses. When plotted by a cartographer, this network serves as a skeletal base onto which outlines of the walls of passages and internal detail are added.

The equipment for surveying is simplistic and commonly consists of a finely calibrated and sightable compass (such as Brunton or Suunto models), a clinometer or hand-level, a sturdy measuring tape of steel or fiberglass, and a notebook. Logistics of traversing difficult, small, or awkward passages precludes the use of larger, more accurate surveying instruments such as theodolites, transits, allidades, stadia rods, and the like. Nonetheless, the “compass-and-tape” system is very efficient; it is both rapid and accurate. An experienced party can easily survey 300 to 1000 m of typical passages in six hours. When the compass is sturdily held or mounted, azimuth readings are accurate to 0.5 degree of arc. Errors of sighting tend to cancel each other rather than accumulate as the survey progresses. Errors in positions of passages are usually less than 0.5 percent; however, the relatively few large errors that occur are often easily detected and corrected (Bassham, 1980; Yannone, 1982). A mathematical treatment of the accuracy of surveying caves is given by Schwinge (1962).

Cartography

Data from surveying can be easily plotted by hand using a protractor and graduated scale. This procedure is tedious and prone to mechanical errors. However, errors can be minimized by first converting original station-to-station measurements (in spherical coordinate form) to latitude and departure (cartesian coordinate form) (Kunath, 1967; Bassham, 1971; Nieland, 1978). Stations are plotted on cartesian graph-paper, survey lines are drawn between stations, and outlines of walls and internal features are added (Rennie, 1965; Irwin, 1970; Nieland, 1975).

Conversion of coordinates is easily accomplished with calculators or computers. Some advanced computational programs calculate errors of closure in surveyed loops and automatically correct the original data by distributing these errors evenly along the loops (Hardy, 1970; Plantz and Schmidt, 1970; Schmidt and Schelleng, 1970; Wilcock, 1970; Rutherford and Amundson, 1974; McKenzie, 1977). Through the use of computer-graphics, many programs automatically plot lines of survey in plan or profile, enabling rapid and accurate drafting of the maps.

MAPPING OF CAVES IN THE EDWARDS PLATEAU

Surveying and cartography during this study were accomplished in the manner outlined above. Most maps presented herein are of grade 5C or 5D according to the British Cave Research Association system of grading for accuracy and detail (Tables 18 and 19).

Symbols used on maps of caves in this study are those in standard use in the United States (Figure 173). Additional and special symbols, where needed, are explained on the individual maps.

Clinometers, such as those of Brunton and Suunto compasses, are normally accurate to only about one or two percent. In some caves geologically subtle features, such as gently dipping strata or passages with low gradients, are difficult to measure with clinometers. Debenham (1964), Palmer (1970), and St. Pierre (1970) describe the “level-tube” method of vertical surveying. A flexible, vinyl tube, open at both ends, is nearly filled with water. Each end of the tube is held upright at a survey station. The water level at each end is the same, enabling a datum of elevation to be transferred from one station to the next. Vertical positions of ceilings, floors, and geologic horizons can be measured relative to this datum. With care, vertical errors can be held to less than 0.1 percent with this method.

Where flow is tranquil, unbroken water-surfaces of streams and pools in caves can be used effectively as horizontal data-planes from which vertical measurements can be made. Long segments of partially flooded caves such as Cave Without a Name, Kendall County, and Neel Cave, Menard County, were mapped in this manner. Changes in levels of pools and streams across cascades and waterfalls were carefully measured where necessary.



Figure 173: Standard symbols used on maps of caves. Compiled and modified from Russell and Raines (1967, p. 109-110), Vineyard (1967), Railton (1969, p. 16-20), Freeman (1975, p. 79), Kastning (1975, p. 165), Nieland (1978, p. 8), Fabre and Audetat (1978), Hedges and others (1979), Daunt-Mergens (1981, p. 95), Sprouse (1982)

Table 18: System of Grading the Accuracy of Survey-Lines[a]

Grade[b]	Definition
1	A sketch of low accuracy made without measurements. Generally includes sketches based on memory.
2	A sketch that is intermediate in accuracy between grade 1 and grade 3. Generally includes sketches made inside a cave but without the use of instruments.[c]
3	A rough, magnetic survey. Horizontal and vertical angles ar emeasured to + or – 2.5°, distances are measure to + or – 50 cm, and errors of station–position are less than + or – 50 cm.
4	A survey that fails to attain all of the requirements of grade 5 but is more accurate than a grade 3 survey.[c]
5	A magnetic survey. Horizontal and vertical angles are accurate to + or –1.0°, distances are accurate to + or –10 cm, and errors of station–position are less than + or – 10 cm. Instruments must be properly calibrated.
6	A magnetic survey that is more accurate than grade 5. Horizontal and vertical angles are generally accurate to + or – 0.5°, distances are accurate to + or – 2.5 cm, and errors of station–position are less than + or – 2.5 cm. The use of tripods or similar techniques are generally required.
X	A survey that is based primarily on the use of a theodolite instead of a compass. A grade x survey must include on the drawing notes on the type of instruments and techniques used, together with an estimate of the probable accuracy of the survey.

[a]Based on Ellis, 1976a, table 1, p. 2.
[b]Grade number is generally followed by a letter indicating accuracy and level of detail in the notetaking (Table 19).
[c]Grades 2 and 4 are for use only when, at some stage of the survey, physical conditions have prevented the surveyor from attaining all of the requirements for the next higher grade and it is not practical to survey again.

Table 19: Classification of Detail in a Survey[a]

Class[b]	Definition
A	All details are based on memory.
B	Details of passages are estimated and recorded in the cave.
C	Measurements of detail are made at survey-stations only.
D	Measurements of detail are made at survey-stations and wherever necessary between stations to show significant changes in shape, size, direction, etc. of passages.

[a]Based on Ellis, 1976a, Table 2, p. 3.
[b]Letter of class is generally affixed to the grad number (Table 18). The accuracy of detail should be appropriate to the accuracy fo the survey line (Table 18) and it is recommended that the following combinations of grades and class be used: 1A; 3B or 3C; 5C or 5D; 6D; XB, XC, or XD.

APPENDIX E
EVALUATION OF REMOTELY SENSED IMAGERY IN MAPPING FRACTURES IN A CARBONATE TERRANE

INTRODUCTION

The Bat Cave Quadrangle, Comal and Bexar Counties, provides an excellent area for evaluating the use of remotely sensed imagery of various types and scales in mapping fractures. *En echelon* faulting along the Balcones Escarpment has created primary sets of fractures parallel to fault-planes and secondary sets at oblique angles to fault-planes. Fractures are numerous, are of different lengths, and have affected landforms to differing degrees. It was expected, therefore, that this comparative analysis would show the level to which fractures could be identified on different types of imagery. Advantages, limitations, and degree of utility of each type of imagery as a viable means of mapping fractures are discussed below.

LINEAMENTS AND FRACTURES

In areas of relatively flat-lying, undeformed, bedded rocks, geologic structures are often observable in aerial imagery as traces or lines, commonly referred to as "lineaments." O'Leary and others (1976, p. 1467) define a lineament as "a mappable, simple or composite linear feature of a surface, whose parts are aligned in a rectangular or slightly curvilinear relationship and which differs distinctly from the patterns of adjacent features and presumably reflects a subsurface phenomenon." Definitions of related terms are given by O'Leary and others (1976) and El-Etr (1976). Care must be exercised in interpreting linear features in areas which have been modified by man. Fence lines, transmission lines, pipelines, boundaries of cultivated fields, and the like are commonly visible as lineations in remotely sensed imagery, and must be discerned as such in order to isolate them from natural lineaments corresponding to geologic structure. Indeed, it is possible for an overzealous interpreter to map more photolineaments than actually exist (Wise, 1976, 1982), especially if he searches for them where he wishes to find them.

Lineaments were first analyzed in the early nineteenth century to describe structural features (Hodgson, 1976) and the method has since progressed to become a useful, geologic tool, particularly through the use of aerial photographs to map joints, faults, and other fractures (Brown, 1961; Hough, 1959, 1960; Isachsen and McKendree, 1977; Kaiser, 1950; Lattman, 1958; Lattman and Matzke, 1961, 1962; Lattman and Nickelsen, 1958; Lattman and Segovia, 1961; Matzke, 1961; Trainer, 1967; Weber, 1972; Wise, 1969; Wobber, 1967). The technique is highly applicable to the search for economic resources including hydrocarbons, ores, and minerals (Blanchet, 1957; Gathright, 1981; Henderson, 1960; Kupsch and Wild, 1958). However, the widest use of photolineaments has been as a means of prospecting for groundwater (anonymous, 1978; Setzer, 1966), with particular emphasis in hydrogeological investigations of carbonate terranes (Hine, 1970; LaRicca and Rauch, 1977; Lattman and Parisek, 1964; Matzke, 1961; Meisler, 1963; Moore and others, 1977; Rauch and LaRicca, 1978; Siddiqui and Parisek, 1971; Sonderegger, 1970; Trainer and Ellison, 1967).

Fractures commonly develop as systematic sets in response to regional tectonism, even in very slightly deformed, sedimentary rocks (Melton, 1929, 1930, 1931; Nelson, 1979; Nickelsen and Hough, 1967, 1969; Parker, 1942, 1969; Rich, 1928). Orientations of systematic fractures may be statistically analyzed and correlated with regional structure and tectonism (Pincus, 1951; Karcz, 1978; Wise and McCrory, 1982).

Joints and faults greatly influence the flow of water through karstic terrane and thereby typically determine orientations of passages in caves (Anderson, 1961; Barnholtz, 1961; Deike, 1969; Kastning, 1975, 1977c, 1981b, 1983b, 1984; Powell, 1975b, 1977). Within the past fifteen years remotely sensed imagery has been increasingly employed in identifying lineaments and mapping and interpreting karst in many states including Alabama (Warren and Wielchowsky, 1973; Wilson, 1977), Arkansas (Barlow and Ogden, 1980, 1982), Kentucky (Kastning, 1983b, 1984; Kastning and Kastning, 1981), Tennessee (Garza, 1971), Texas (see discussion below), and West Virginia (Ogden, 1974, 1977; Ogden and Reger, 1977). Both conventional and infrared imagery has been used in studies of karstic systems (Brown, 1972a,b; Rinker, 1975).

STUDIES OF LINEAMENTS IN TEXAS

Remote sensing has been used for some time to investigate fractures in the Edwards Plateau and adjacent regions of Texas (Barton, 1933; Boyer, 1968; Boyer and others, 1961; Cuyler, 1931; Dix and Jackson, 1981; Finley and Gustavson, 1981; Lattman and Olive, 1955; Melton, 1934). McQueen (1963) and Boyer and McQueen (1964) mapped faults and joints in the field for an area of about 480 km^2 in southern Burnet County and compared these to photolineaments in the same area, concluding that photolineaments are largely representative of fractures in the rocks. Fractures were mapped for the southern Edwards Plateau (Wermund, 1976; Wermund and Cepeda, 1977a,b; Wermund, Cepeda, and Bell, 1974; Wermund and others, 1978) and for part of Val Verde County (Leonard, 1977a,b, 1978) as part of a regional analysis of tectonics and recharge and flow of groundwater. As part of a regional geologic study Freeman (1964a,b,c, 1965, 1968), Sharps (1963, 1964), and Sharps and Freeman (1965) mapped photolineaments in Val Verde, Terrell, and Brewster Counties of the southwest Edwards Plateau. Webster (1978, 1980, 1982) mapped photolineaments in Edwards, Kinney, and Val Verde Counties as part of his structural analysis of the Carta Valley Fault Zone. Caran and others (1981) have inferred geologic structure along the Balcones-Ouachita trend based on mapping of photolineaments and Cannon (1974b,c) applied radar-imagery to mapping of environmental geology for a part of the Edwards Plateau.

Much of the present study concerns the influence of geologic structure on karstic processes in central Texas. Some preliminary findings have been published during the course of the investigations (Kastning, 1977c, 1980b,e, 1981b, 1983a,b,c, 1984).

ANALYSIS OF IMAGERY

Six sets of photographic imagery were used to map fractures in the Bat Cave Quadrangle area, Comal and Bexar Counties (Chapter 10 and Table 20). These are evaluated here in order of decreasing flight altitude.

LANDSAT (ERTS) Band 5 Imagery

The one satellite-photograph used in this analysis was a glossy 15 cm by 15 cm print of a LANDSAT (ERTS) Band 5 image (Grimshaw, 1976, frontispiece). It covers an area of approximately 8350 km^2 and includes the cities of Austin, San Marcos, New Braunfels, and the northeastern fringe of greater San Antonio. At this very small scale (1:600,000), it does not provide enough detail to map landforms or many fractures. However, major faults of the Balcones Fault Zone are easily identifiable on this image. Down-faulted blocks of carbonate rocks stand out as darker shades of gray in contrast to the lighter shades for surfaces of limestone on the Edwards Plateau and for the post-Cretaceous rocks southeast of the Comal Springs Fault.

This photograph confirmed the position of the five major faults mapped by Newcomb (1971, plate 1) and shown in Figure 115. It was the only type of imagery which clearly showed expression of each major fault. This supports the general observations that (1) regionally extensive features are less likely to be detected in photography at close range, and (2) imagery from high altitudes is more applicable in identifying large features. Orbital and suborbital, remote sensing is a valuable tool for regional analysis and has been successfully used in central Texas in hydrogeomorphic analysis of potential for flooding (Baker and others, 1974; Baker, Holz, Hulke, and others, 1975; Baker, Holz, and Patton, 1975; Baker and Patton, 1975; Baker and Holz, 1978). However, imagery at this scale is very limited for use in detailed mapping of areas the size of a topographic quadrangle or smaller.

Color-Infrared Imagery

Color-infrared transparencies of two scales (1:122,000 and 1:62,100) used in this study were valuable in mapping units of landforms (e.g. deeply or moderately dissected karstic terrane, karstic valleys, terraces) but were less satisfactory in

Table 20: Imagery Used in Mapping Procedures, Bat Cave Quadrangle, Comal and Bexas Counties, Texas.

Type of Photography	Scale (calculated from Photograph)	Date Flown	Roll or Composite Number	Frame or Registration Numbers	Agency or Firm Flown by	Source of Loaned Photographs
Black and White (Panchromatic) Stereo coverage (60% overlap)	1:77,400	2-9-67	6185-4	34-36	Tobin Surveys, Inc. San Antonio, Texas	Texas Water Development Board (TWDB) (through Bureau of Economic Geology [BEG])
Black and White (Panchromatic) Controlled photo-mosaic	1:25,600	Prior to 1964	3,S-14E-3	017H	Tobin Surveys, Inc. San Antonio, Texas	BEG
Black and White (Panchromatic) Stereo coverage (60% overlap)	1:73,300	12-12-52	VV AJ M 6	785-787	U. S. Army Map Service	BEG
Color Infrared	1:62,100	1-31-74	36	131	National Aeronautics and Space Administration (NASA)	TWDB (through BEG)
Color Infrared	1:122,000	1-31-74	34	1794	(NASA)	TWDB (through BEG)
Black and White Band 5 LANDSAT (ERTS)	1:599,700	7-21-73	-	-	(NASA)	Frontispiece, Grimshaw (1976)

Table 21: Evaluation of Imagery Used in Mapping Features

Type of Imagery	Scale	Suitablility in Identifying		
		Major Faults	Lineaments	Landforms
LANDSAT Black-and-white	1:599,700	Good	Extremely Poor	Very Poor
Color Infrared	1:122,000	Fair	Fair	Good
Color Infrared	1:62,100	Fair	Poor to Fair	Good
Black-and-white Mosaic	1:25,600	Poor to Fair	Very Good	Good
Black-and-white Stereo	1:73,300	Poor	Excellent	Very Good
Black-and-white Stereo	1:77,400	Poor	Excellent	Very Good

mapping fractures. Variations in hue and chroma in the transparencies facilitated recognition of vegetational assemblages and thereby differences in soils, substrate, and bedrock. This accentuated stratification in the bedrock and aided in identification of landforms. Patterns of vegetation that commonly vary across faults aided in identification of faults. Fractures were more difficult to map on available, color-infrared images than on conventional, black-and-white photographs for four reasons. Firstly, smaller-scale (1:122,000), infrared photographs were nearly twice the scale as that of stereo-pair, black-and-white photographs (1:73,300 and 1:77,400), thereby lessening resolution available to identify subtle lineaments. Secondly, the larger-scale (1:62,100), infrared photographs used were too dark (due to film degradation or processing error), preventing detection of subtle lineaments where a wide range of light-value was necessary for their recognition. Thirdly, flight-lines at the 1:62,100 scale did not overlap, and consequently, the northernmost part of the area was not photographed at this scale. Fourthly, both scales of infrared photography had no more than 20 percent frame-to-frame overlap, thereby eliminating stereoscopic viewing, a method found to be very useful for recognition of fractures.

Black-and-White Photomosaic

A Tobin photomosaic (1:25,600 scale) was available for the Bat Cave quadrangle and proved to be of considerable help in identifying and mapping fractures. This "controlled" mosaic was compiled by photogrammetric adjustment from standard, black-and-white, aerial photography and was processed with contrast to enhance recognition of lineaments (Wermund, Cepeda, and Bell, 1974; Wermund, 1976; Wermund and Cepeda, 1977a,b; Wermund and others, 1978). However, very subtle lineaments, requiring a broader range in light-value for recognition, were difficult, if not impossible, to identify on the photomosaic. Because the mosaic could not be viewed stereoscopically, some fractures were not visible on this imagery.

Black-and White Stereoscopic Imagery

Stereoscopic pairs of black-and-white imagery (1:77,400 and 1:73,300 scales) were significantly better for mapping of landforms and fractures. A moderate range in contrast and light-value allowed good resolution of linear features. Stereoscopic viewing accentuated relief and enhanced detection of alignments of valleys and small scarps, both of which aid in recognition of lineaments. Landforms were best identified by stereoscopic viewing because they are three-dimensional features. However, very small landforms, such as small sinkholes (less than 5 m in diameter), cave entrances, and the like were not visible at this resolution.

SUMMARY

Mapping of landforms and patterns of fractures in the carbonate terrane of the Edwards Plateau was facilitated by use of aerial photography. By using imagery of various types and scales it was possible to construct a map of fractures (Chapter 10, Figure 132).

Types of imagery are rated according to their suitability at different levels of recognition of fractures (Table 21). This comparison indicates that:

1. Low-altitude, large-scale (1:75,000) stereo-photographs provide the greatest resolution for recognizing and mapping lineaments.

2. Very-high-altitude, orbital or suborbital, small-scale (1:600,000) imagery is most useful in identifying regional patterns such as major faults.

3. High-altitude, small-scale (1:122,000 to 1:62,100) color-infrared photographs are suitable for recognizing some major lineaments, but do not offer sufficient resolution for mapping most, smaller lineaments. If such imagery were available in stereo-pair format, recognition of fractures would be greatly improved, as would be the discrimination of assemblages of vegetation, soils, and substrate. This imagery is ideally suited for mapping units of landforms.

4. Controlled, photomosaic sheets, produced with high contrast, provide good resolution of landforms and photolineaments; however, this imagery is not as useful as low-altitude, black-and-white, stereoscopic imagery. Photomosaics should not be used exclusively if it is desirable to produce the most complete map of fractures by remote sensing, as many additional lineaments are visible in standard (1:75,000 to 1:20,000), stereo-pair photographs of the same area. However, controlled mosaics do offer one advantage; most have been photogrammetrically assembled and show features in correct map projection.

5. The best method of mapping structural features in an area is to use imagery of different types and scale and to combine the results into a single map, as was done for the Bat Cave quadrangle in Chapter 10 (Figure 132).

APPENDIX F
DRAINAGE DENSITY OF CARBONATE TERRANE

INTRODUCTION

Quantitative measures of the geometric, spatial, and distributive aspects of the landscape, collectively known as morphometric variables, often provide insight toward resolving the geomorphic history of an area (Chorley, 1972; Clark, 1966; Doornkamp and King, 1971; Slaymaker, 1966). Numerous parameters have been devised and proposed (Horton, 1945; Strahler, 1957, 1964, 1968). Early examples of morphometric analysis pertained to streams and fluvial networks; however, more recently the technique has been extended to other systems including karst (Baker, 1973; Kastning and Kastning, 1980; Kemmerly, 1976; LaValle, 1965, 1967, 1968; McConnell and Horn, 1972; Ogden and Reger, 1977; Palmquist, 1977; E.L. White, 1975a, b, 1976, 1977; Williams, 1966a, b, 1971, 1972a, b).

Drainage density is the sum of the lengths of all fluvial channels within a boundary divided by the area enclosed by that boundary (Horton, 1932). Drainage density is expressed as

$$D_d = L/A, \qquad \text{(F-1)}$$

where L is the total length of streams and A is the area, both in the same system of units (Horton, 1945). Drainage density characterizes the degree of development of drainage within an area, and it is one of the most useful variables of morphometric analysis. It has been applied in studies of (1) the dynamics of discharge within fluvial basins (Langbein and others, 1947; Maxwell, 1960; Carlston, 1962, 1963; Howe and others, 1966, 1967; Gregory and Walling, 1968; Rodda, 1969; Trainer, 1969; Patton, 1975; Patton and Baker, 1976; Orsborn, 1976; Graf, 1977; Murphy and others, 1977), (2) the classification of drainage-basins (Woodruff, 1964; Eyles, 1971), (3) the effect of lithology on the characteristics of drainage-basins ((Melton, 1957; Orsborn, 1970; Gardiner, 1971, 1972; Wilson, 1971; Gregory and Gardiner, 1975; Skolasky, 1978), (4) sediments yielded from drainage-basins (Stall and Bartelli, 1959; Abrahams, 1972), (5) the development of erosional topography (Horton, 1945; Smith, 1950; Schumm, 1956; Melton, 1957; Morisawa, 1962; Chorley and Morgan, 1962; Strahler, 1964, 1968; Gregory, 1966a; Anderson and Lohnes, 1969; Wilson, 1971; Gardiner and others, 1977; Gardiner, 1978), (6) the relationships between fluvial networks and climate (Melton, 1957; Chorley, 1957; Cotton, 1964; Carlston, 1966; Gregory and Gardiner, 1975), and (7) the evaluation and delimitation of natural regions (Gardiner, 1974a, 1976).

Drainage density can vary considerably from one area to another in response to differences in lithology, structure, soil, and climate. It can be correlated with other physiographic and geologic variables in order to determine cause and effect in the evolution of landscape.

AREA OF STUDY

A study of the relationship between development of drainage and the geologic setting was initiated in the southeastern Edwards Plateau. Drainage densities at two localities were measured from topographic maps. The first (Area 1) included six 7.5-minute quadrangles in eastern Real County and western Banderas County. The second (Area 2) included twelve 7.5-minute quadrangles in southeastern Gillespie County, western Blanco County, eastern Kendall County, and western Comal County. These areas are underlain by the Glen Rose Formation and carbonates of the Edwards Group. The two localities were chosen because they have been deeply dissected by surficial drainage in response to heightened relief imposed by faulting along the Balcones trend.

METHODS OF ANALYZING DRAINAGE DENSITY

Computation of values of drainage density is subject to judgement and to the procedure used in measuring total length of streams in a given area. Morisawa (1957), Werrity (1972), Chorley and Dale (1972), Drummond (1974), Smart (1978), and Kennedy (1978) have addressed the problem of accuracy in determining length of streams. There are several ways this can be done. Streams may be measured directly in the field; however, even though this is highly accurate, it is generally impractical owing to the great time and effort involved. Secondly, lengths of streams may be measured from aerial photographs. The accuracy obtained is a function of scale and clarity of the photograph. Hulke (1978) found that interpretation of stereo-pair, aerial photographs of low-altitude (1:14,500 scale) gave a high resolution of streams, slightly higher than that obtained from topographic maps (1:24,000 scale). Coffman (1970) asserted that aerial photographs are much more accurate than topographic maps at 1:24,000 scale for mapping drainage. Thirdly, lengths can be interpreted from topographic maps.

Blue lines on these maps indicate water-courses, and measuring their length can approximate length of streams. However, Morisawa (1957) has shown by statistical analysis that the "blue-line" method results in the omission of many streams of lower order and gives an unrealistically low value of stream-length, and therefore, it should not be used. A more accurate method is to supplement the blue lines on available maps by drawing in lines of drainage indicated by crenulations in the contours (contour "V's"); that is, extending all lines of drainage up each small valley or channel, to a point where the crenulations no longer indicate existence of the channel (Figure 174). Morisawa (1957) showed that there is no significant difference in accuracy between the "contour-crenulation" and "field-survey" methods. The contour-crenulation technique is generally recommended as the most efficient method for measuring lengths of streams (Miller, 1953; Morisawa, 1957; Brush, 1961; Smart and Surkan, 1967; McCoy, 1971; Gregory and Walling, 1973; Gardiner, 1974c).

A less important measure of erosional texture of the landscape is the texture ratio, which is the number of crenulations in the contour within the basin having the maximum number of crenulations, divided by the perimeter of the basin (Smith, 1950). Texture ratio and drainage density are linearly related on a log-log basis (Strahler, 1957). This fact supports the use of the contour-crenulation technique.

An inconsistency arises in the case where a sinuous stream occupies a floodplain in the bottom of a valley. The total stream-length, L_s, increases with sinuousity, thereby inflating drainage density. However, the total length of *valleys* is a better measure of L_s, because it better characterizes the average length of the route of flow. Because "length of overland flow is one of the most important independent variables affecting both the hydrologic and physiographic development of drainage basins" (Horton, 1945, p. 284), and it better reflects the efficiency of drainage, Dingman (1978) proposed that total valley-length (as used by Schumm, 1963) is a more appropriate variable than total stream-length in the definition of drainage density. However, care must be exercised where dry valleys exist, because these valleys may not necessarily be indicative of present-day geomorphic processes (Gregory, 1971).

Resolution of stream channels depends upon the scale of aerial photographs or maps used (Schneider, 1961; Giusti and Schneider, 1962; Gregory, 1966b; Gregory and Walling, 1968; Smart, 1968; Haggett and Chorley, 1969, p. 75; Werrity, 1972; Pethick, 1975; Gardiner, 1977). Obviously large-scale photographs or maps yield a better approximation of true stream-length than do those of small scale. Small channels and rills not indicated on standard, 1:24,000-scale, topographic maps may be visible in aerial photography of low-altitude, as shown for the Edwards Plateau by Hulke (1978).

Direct measurement of stream-length from topographic maps (1:24,000 scale) is tedious and somewhat imprecise as it is necessary to roll a measuring wheel over all indicated streams. Fortunately, reasonably accurate and much more rapid techniques have been developed that provide a very reliable estimate of stream-length and thereby drainage density (Carlston and Langbein, 1960; Wilgat, 1966; McCoy, 1969, 1970, 1971; El-Ashry, 1971; Coffman and others, 1971; Donahue, 1974; Van Asch and Von Steijn, 1973; Mark, 1974, 1975; Karcz, 1978; Richards, 1979; Gardiner, 1974b, 1979). Carlston and Langbein (1960) developed the "line-intersection" method for measurement of drainage density. The procedure is as follows:

1. A map of drainage is constructed wherein streams are traced onto topographic maps by using blue lines printed on the maps and extending lines to the maximum

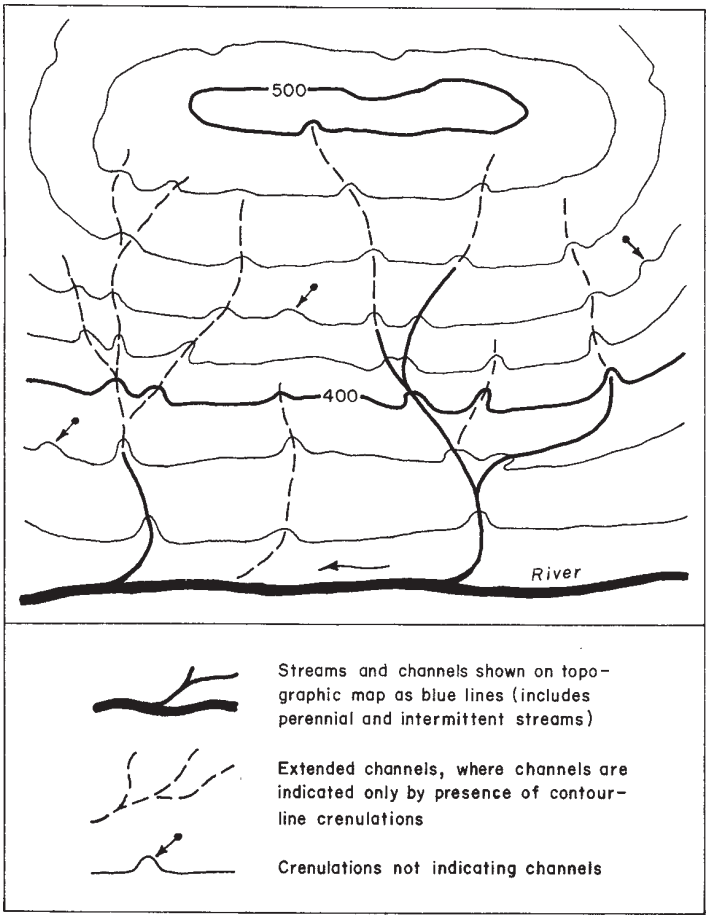


Figure 174: Method of extending lines of drainage on topographic maps. Existing lines of drainage are extended headwardly through crenulations in contours. Modified from Gardiner (1974c) and Skolasky (1978, p. 23).

headward limit of crenulations in contours (Figure 175A and B).

2. An overlay, or transparent template, is constructed upon which a random pattern of lines is drawn (Figure 175C). A random pattern is chosen to compensate for any strong, directional orientations of fluvial networks. Directional bias imposed on this method by lithology or structure within the basin is minimized or eliminated.

3. The transparent template is placed over the map of the drainage (Figure 175D).

4. The total number of intersections (N) of traverse lines on the template (Figure 175C) with extended streams (Figure 175B) are counted for the area as shown in Figure 175D, where $N = 38$.

5. The total length of lines (L) on the template falling within the area is measured.

6. Drainage density is computed from the number of intersections using an empirical equation.

Carlston and Langbein (1960) theorized by geometrical construction that drainage density (D_d) can be computed from the number of intersections (N) and total line-length (L) using the linear relationship

$$D_d = 1.41(N/L), \qquad \text{(F-2)}$$

where D_d and L are in the same system of units (for example, kilometers or miles). McCoy (1971) empirically measured drainage density for 22 basins in the United States and derived the regression-relation.

$$D_d = 1.12 + 1.27(N/L),\ r^2 = 0.86, \qquad \text{(F-3)}$$

where D_d is in km^{-1} and L is in km. Mark (1974), however, re-analyzed the geometrical derivation of Carlston and Langbein's equation and determined that the correct relationship should be

$$D_d = 1.571(N/L) \qquad \text{(F-4)}$$

Performing a statistical analysis of McCoy's 22 basins plus 48 others, Mark verified that the ratio of drainage density to N/L should be 1.571 and he found equation F-4 to agree with empirical data with less than 10% root-mean-square deviation. Mark also determined that the line-intersection method of determining drainage density takes approximately one half the time as does the measuring-wheel method.

The drainage-basin has been acknowledged to be the fundamental geometrical component of the landscape (Chorley, 1969). Determinations of drainage density have routinely been made using the drainage-basin as the areal component, and the approach involved delineating the boundaries of the watershed and network of streams, and measuring lengths of streams and areas of basins. For this study it was decided that this method was unduly laborious. Instead, a grid was constructed over the area under study and each square of the grid (quadrat) was analyzed as in steps 1 through 6 above (Figure 175, A through D, represents a single quadrat). The grid method is similar to that of Barnes (1964), Gardiner (1971, 1979), Gregory and Gardiner (1975), Dacey and Krumbein (1976), and Singh (1976). A quadrat 1.609 km by 1.609 km (1 mile square) was chosen.

The template of lines used in this study (Figure 175C) was "calibrated" by actually measuring total stream-length for 27 quadrats using a map-measuring wheel, determining intersections of lines for the same quadrats, and constructing a scatter diagram (Figure 176). A regression-line fit gives

$$D_d = -0.12 + 1.56\ (N/L),\ r^2 = 0.93 \qquad \text{(F-5)}$$

This relationship compares favorably with the empirical relation of McCoy (1971) and the theoretical relationship, $D_d = 1.41\ (N/L)$, of Carlston and Langbein (1960) (Figure 176). However, equation F-5 is nearly identical with, and thereby corroborates, Mark's (1974) relation (equation F-4). Values predicted by these two relations are within 0.2 km^{-1} of each other (Figure 176). Therefore Mark's relationship, $D_d = 1.571(N/L)$, was used to calculate drainage densities for selected areas on the Edwards Plateau.

Drainage density has been used as a predictor of characteristics of a stream-hydrograph, such as the height of flood-peaks and the length of the recession-curve. It has also served as an indicator of the ratio of direct runoff to baseflow, and a measure of rate of infiltration and average runoff. However, Dingman (1978) cautions against direct application of drainage density as a predictive parameter because these relationships vary with climate, storage in the channel and floodplain, and degree of saturation of soil. Nevertheless, measured drainage densities may be compared to indicate relative differences in streamflow, infiltration, and baseflow among areas of different topographic, geologic, or climatic settings

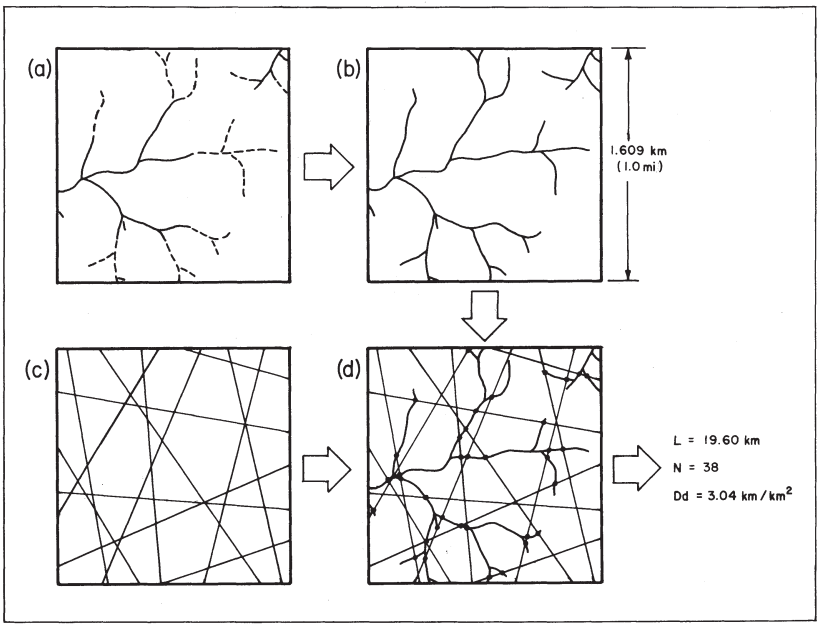


Figure 175: Line-intersection method of measuring drainage density. (A) Lines of perennial and intermittent drainage (solid lines) indicated by blue lines on topographic map are extended (dashed lines) as explained in Figure 174. (B) Extended fluvial network. (C) Configuration of randomly drawn lines on transparent template-overlay. (D) Superposition of template (C) on extended lines of drainage (B). Nodes indicate intersections. Values given here correspond to the example depicted and are calculated as explained in the text.

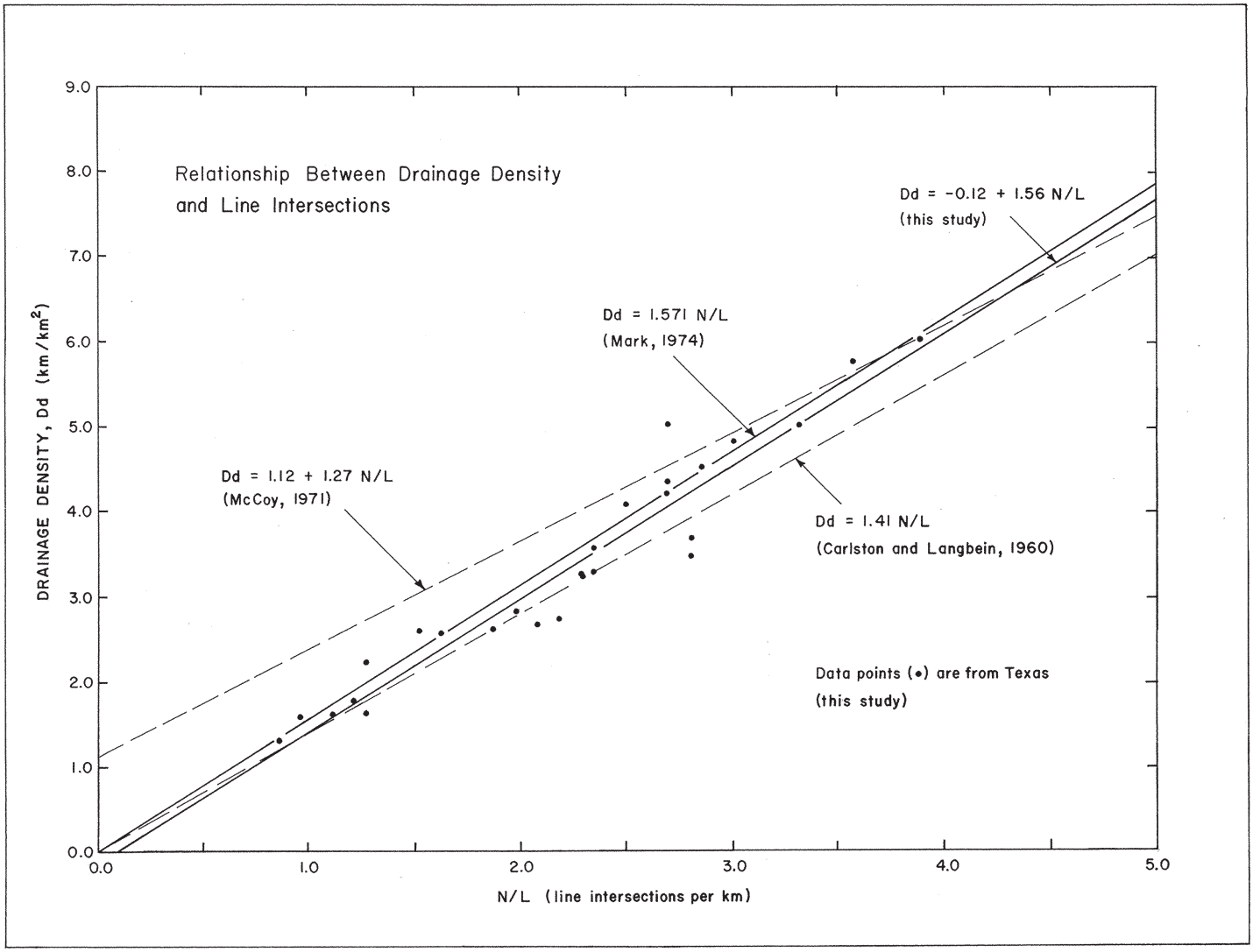


Figure 176: Relationship between drainage density and line-intersections. Empirical results of previous investigations and this study are compared.

APPLICATION TO SOUTHEASTERN EDWARDS PLATEAU

Maps of drainage density of Areas 1 and 2 are shown in Figure 177. Zones of varying intensity of drainage density are readily apparent. Comparisons of these maps with geologic maps (Barnes, 1974b, 1981a) shows the following relationships:

1. Drainage density is high (7 to 11 km per km^2) where the Fort Terrett Formation forms a resistant cap along drainage-divides or where the upper member of the Glen Rose Formation is exposed just beyond the limit of the caprock. This is more pronounced in Area 2 than in Area 1.

2. Drainage density is moderate (3 to 7 km per km^2) on the remainder of the Glen Rose Formation in both areas and on much of the Devils River Formation in Area 1. Drainage density over the Hensell Sand is also moderate (3 to 7 km per km^2).

3. Low values of drainage density (less than 5 km per km^2) are correlative with either alluvial deposits in valleys of larger streams (e.g. Frio and Sabinal Rivers in Area 1 and Pedernales, Blanco, and Guadalupe Rivers in Area 2) or with outcrops of pre-Cretaceous rocks (in the northern part of Area 2).

4. Drainage density is generally higher where slopes in the topography are steepest, corresponding to those areas most dissected by fluvial erosion. Topographic slopes are steeper on the Fort Terrett Formation, especially at the boundaries of its outcrop where steep cliffs are upheld by the Fort Terrett and the underlying Comanche Peak and Walnut Formations in Area 2. Slopes developed on the Glen Rose Formation, although the formation develops a characteristic "stair-step" topography, are shallower over the extent of the outcrops.

The correlations observed among drainage density, stratigraphic units, and slope are consistent with the physical properties of the rocks. Within a particular region rates of weathering and development of slopes depend on the hardness of the bedrock, thickness of the beds, density of fractures, and stability of slopes. High values of drainage density indicate a greater ease and rapidity by which streams incise into the landscape, and this in turn is manifested by the development of steeper slopes. Incision is typically greatest on relatively impervious ground where surficial runoff exceeds the rate of infiltration (Langbein and others, 1947).

A gently rolling topography has formed on the Hensell Sand and "stair steps" or benches of low relief are developed where the lithology varies among successive beds. The formation varies from conglomeratic material to silty clay, but much of it is sandy and dolomitic (Stricklin and others, 1971). As a result incisions of streams and drainage densities are moderate.

The upper member of the Glen Rose Formation develops slopes of moderate steepness and drainage densities of intermediate values because it consists of an alternating sequence of thin, resistant beds of dolostone, mudstone, limestone, and non-calcareous clay and marl. Moreover stability of slopes developed on the Glen Rose is moderate to high. "Stair steps" developed on the sequence are low in relief over the distances used to measure drainage density.

The Walnut Formation (Area 2) is generally too thin to be of great consequence in the development of drainage. However, the stability of slopes developed on it is moderate to high and it maintains steep slopes over short vertical distances. It acts in conjunction with the Comanche Peak and Fort Terrett Formations to uphold steep slopes and promote high drainage densities.

The Comanche Peak Formation (Area 2) is a nodular, soft limestone. It weathers to produce moderate to steep slopes. Because it is thin (5 m or less) at this locality, it is also of little consequence by itself in affecting drainage densities.

The Fort Terrett Formation in Area 2 is a hard, massive, nodular, marly limestone and microgranular dolomite containing abundant chert. The upper part is

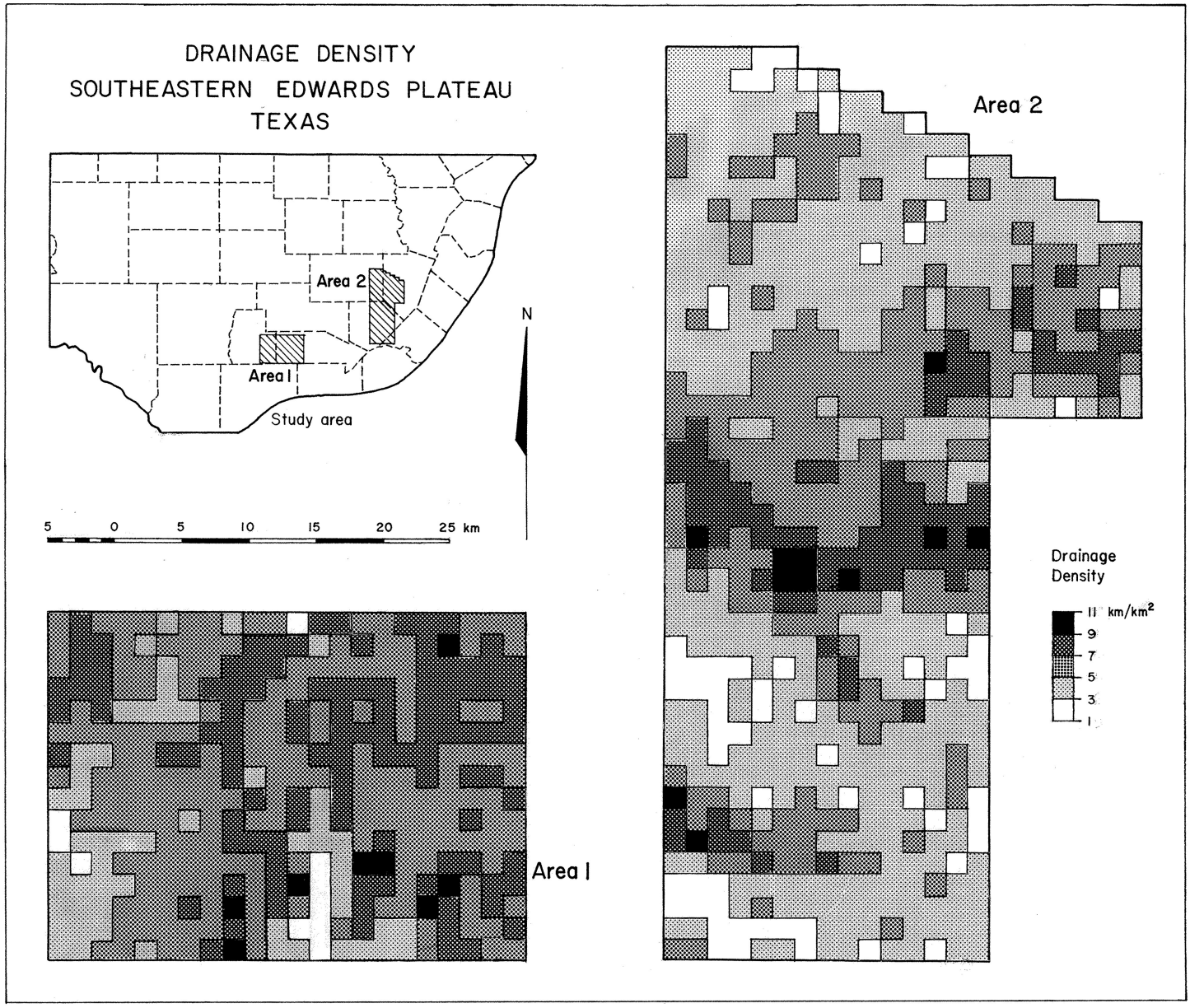


Figure 177: Maps of drainage densities of two areas within the southeastern Edwards Plateau.

bench-forming. Rates of infiltration are low in the fine-grained, dense beds, and they are moderate to high in the coarse-grained, cavernous beds. Additionally, the Fort Terrett has a high stability of slopes. Hardness and locally reduced infiltration-capacity allows this unit to develop a more deeply dissected topography and higher drainage density than those associated with neighboring exposures of the Glen Rose Formation.

In Area 1 the Fort Terrett and Segovia Formations have moderate to high values of drainage density. Locally, the Segovia forms a resistant caprock that protects the underlying formations from erosion and retreat of slopes. The margins of the outcrop of the Segovia and the adjacent exposures of the Fort Terrett are consequently deeply dissected and maintain steep topographic slopes. Both the Fort Terrett and Segovia have high stability of slopes.

The speleogenetic importance of drainage density is that it is a partial measure of karstification and the development of caves. Where slopes are steepest and drainage densities are highest, caves are well preserved under caprock (Area 1). In some areas of moderate drainage density, such as the exposures of the lower Glen Rose Formation, long caves occur near baselevel along major streams. These results are similar to those of a similar study initiated in the Mississippian Plateau of western Kentucky (Kastning and Kastning, 1980).

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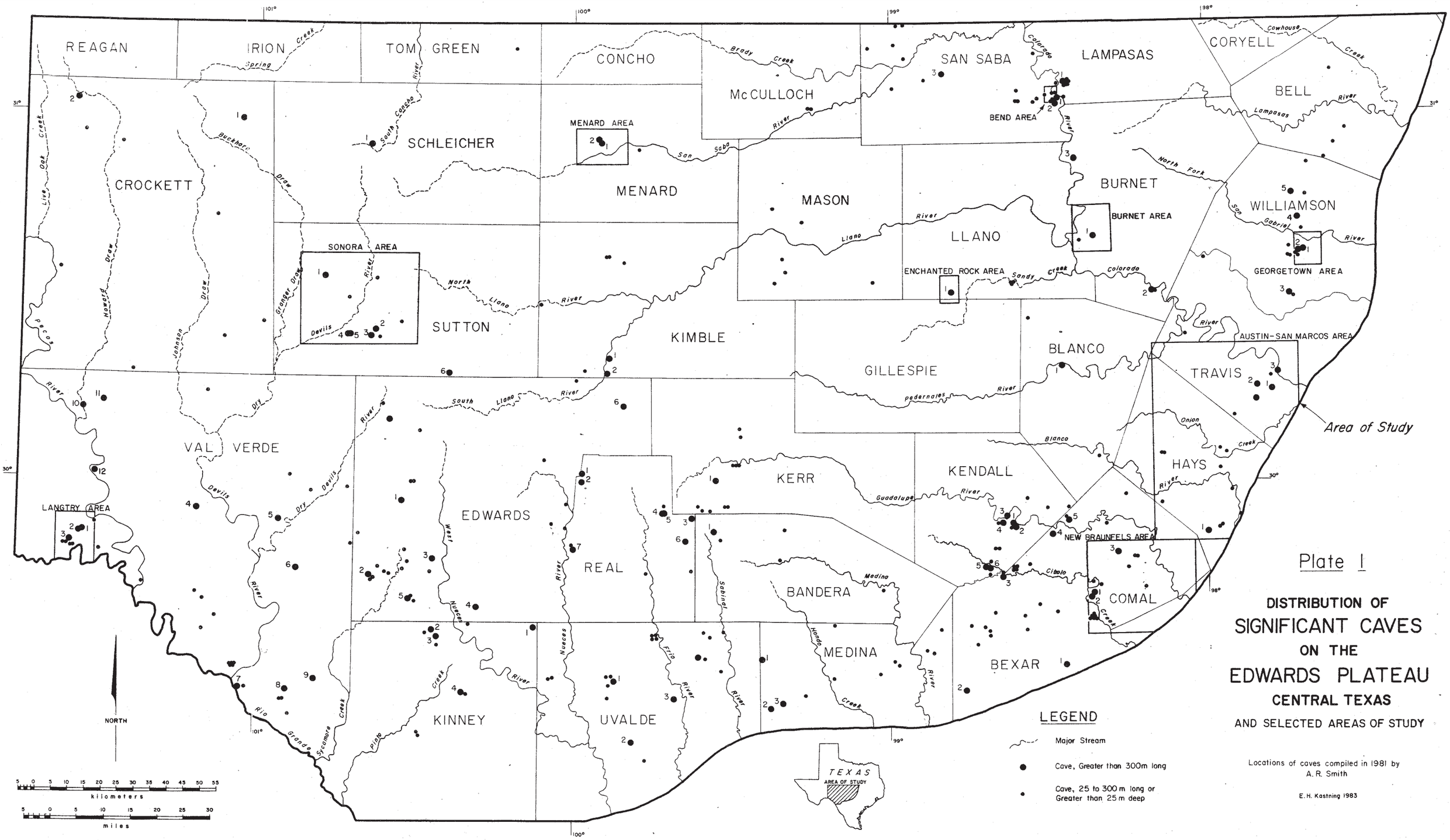


Plate 1

DISTRIBUTION OF SIGNIFICANT CAVES ON THE EDWARDS PLATEAU CENTRAL TEXAS AND SELECTED AREAS OF STUDY

Locations of caves compiled in 1981 by A. R. Smith

E. H. Kastning 1983

Area of Study

NORTH

5 0 5 10 15 20 25 30 35 40 45 50 55
kilometers

5 0 5 10 15 20 25 30
miles



LEGEND

Boundaries of Counties

Perennial Streams

Intermittent Streams

Plate 2

SURFICIAL DRAINAGE ON THE EDWARDS PLATEAU CENTRAL TEXAS

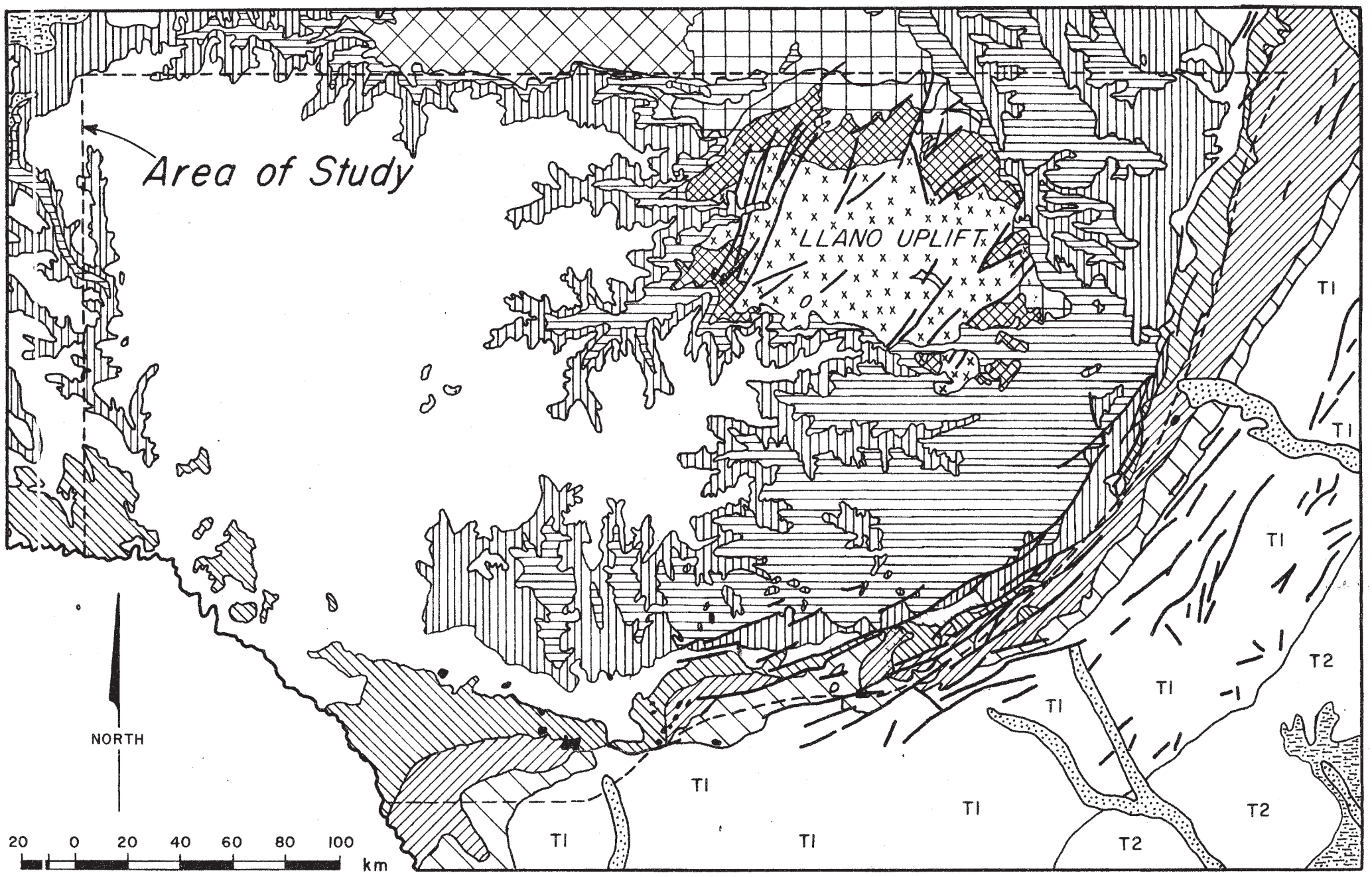
Streams traced from blue lines on U.S. Geological Survey 1:250,000-scale topographic maps (1953-1975).

K. M. Kastning and E. H. Kastning 1983

Plate 3

GENERAL GEOLOGY

EDWARDS PLATEAU, CENTRAL TEXAS



LEGEND

- Fluvial Deposits (Quaternary)
- Pleistocene Deposits (Quaternary)
- T2 — Oligocene, Miocene, and Pliocene Formations (Tertiary)
- T1 — Paleocene and Eocene Formations (Tertiary)
- Navarro Group (Upper Cretaceous)
- Taylor Group (Upper Cretaceous)
- Eagle Ford Group and Austin Chalk (Upper Cretaceous)
- Washita Group (Lower and Upper Cretaceous)
- Fredericksburg Group (Lower Cretaceous)
- Trinity Group (Lower Cretaceous)
- Permian Formations
- Pennsylvanian Formations
- Ordovician through Mississippian Formations
- Precambrian and Cambrian Formations
- Volcanic Rocks (Cretaceous)
- Fault

E. H. Kastning 1983

Plate 4

STRATIGRAPHIC COLUMN

PRECAMBRIAN AND PALEOZOIC ROCKS

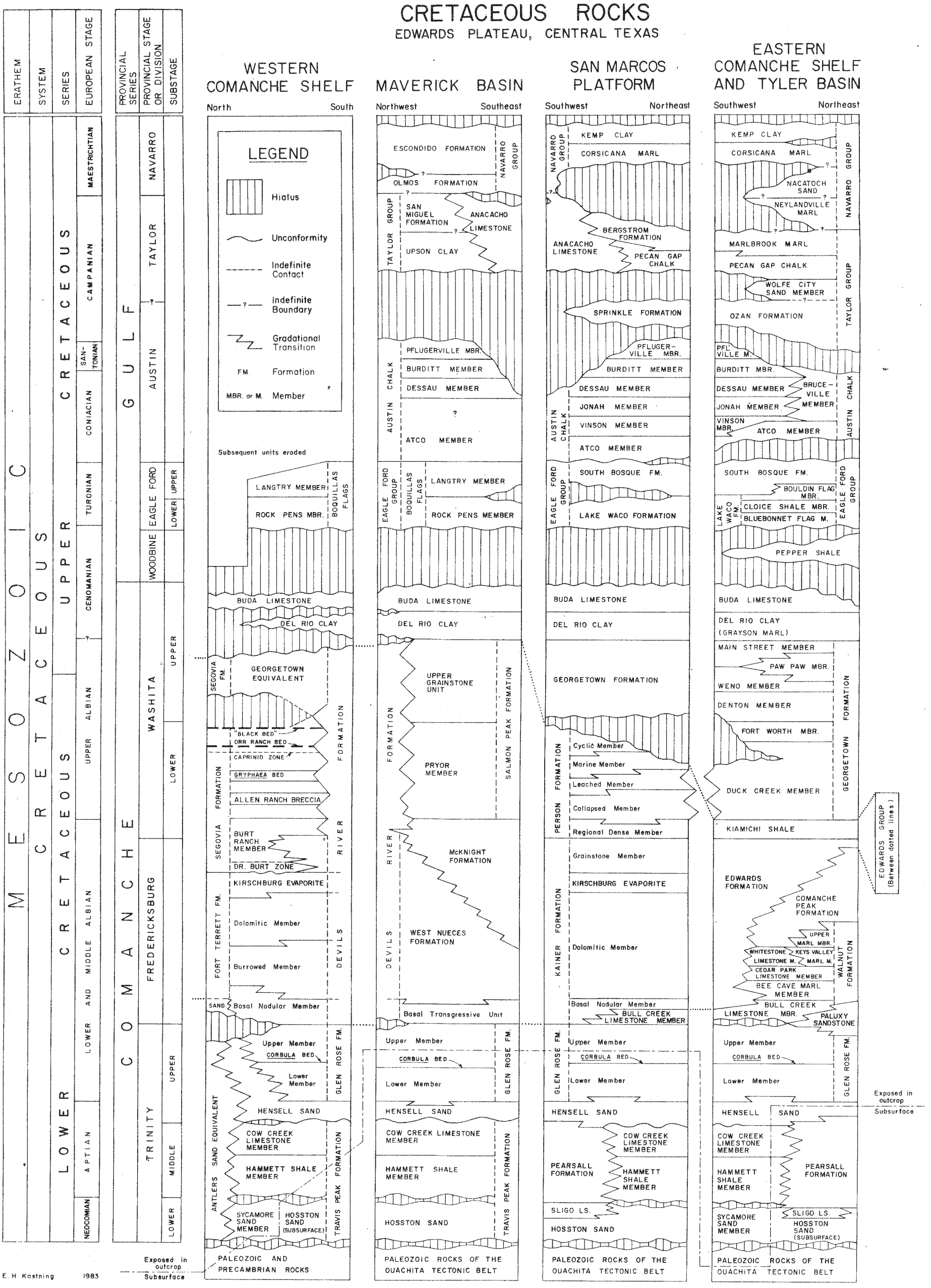
CENTRAL TEXAS

ERATHEM	SYSTEM	SERIES	LITHOLOGY	MEMBER	FORMATION	GROUP
PALEOZOIC	PERMIAN	WOLFCAMP		Includes OVERALL LS. and HORDES CREEK LS.	ADMIRAL FORMATION	WICHITA-ALBANY
				Includes COLEMAN JUNCTION LS.	COLEMAN JUNCTION FM.	
				Includes GOULDBUSK LS., IBEX LS., CAMP COLORADO LS., and unnamed SS.	MORAN FORMATION	CISCO
				Includes STOCKWETHER LS., SADDLE CREEK LS., and unnamed sandstone	PUEBLO FORMATION	
		? ?		Includes BRECKENRIDGE LS. and unnamed sandstone	HARPERSVILLE FM.	
		VIRGIL		Includes SPECK MOUNTAIN LS., IVAN LS., and various unnamed sandstone members	THRIFTY FORMATION	
				Includes GUNSIGHT LS., BUNGER LS., NORTH LEON LS., and unnamed sandstone mbrs.	GRAHAM FORMATION	
		MISSOURI			HOME CREEK LS.	CANYON
					COLONY CREEK SHALE	
					RANGER LIMESTONE	
				Includes CORN CREEK LS. and unnamed sandstone mbr.	PLACID SHALE	
					WINCHELL LIMESTONE	
					CEDARTON SHALE	
					ADAMS BRANCH LS.	
		DES MOINES		Includes various sandstone members	STRAWN FORMATION	STRAWN
		MORROW-? ATOKA		Includes 7 members as mapped by Plummer (1943)	SMITHWICK SHALE	BEND
					MARBLE FALLS LS.	
					Unnamed phosphorite	
	MISSISSIPPIAN	KINDERHOOK-CHESTER		Found in collapse structures	BARNETT FORMATION	
					CHAPPEL LIMESTONE	
				Unnamed phosphorite	HOUY FORMATION	
		? ? UP.		DOUBLEHORN SHALE		
				IVES BRECCIA		
	DEVONIAN	MIDDLE		Found in collapse structures or as fillings of joints or caves	Unnamed limestone	
					ZESCH FORMATION	
					BEAR SPRING FM.	
					STRIBLING FORMATION	
		LOWER			PILLAR BLUFF LS.	
	SIL.				STARCKE LIMESTONE	
	ORDOVICIAN	UPPER			BURNHAM LIMESTONE	
		LOWER			HONEYCUT FORMATION	ELLENBURGER
					GORMAN FORMATION	
				STAENDEBACH MEMBER	TANYARD FORMATION	
				THREADGILL MEMBER		
		? ?		SAN SABA MEMBER	WILBERNS FORMATION	MOORE HOLLOW
				POINT PEAK MEMBER		
	CAMBRIAN	UPPER		MORGAN CREEK LS. MBR.		
				WELGE SANDSTONE MBR.		
		?		LION MOUNTAIN SS. MBR.	RILEY FORMATION	
		MID.		CAP MOUNTAIN LS. MBR.		
				HICKORY SANDSTONE MBR.		
PRECAMBRIAN	MIDDLE PRECAMBRIAN				CLICK FM. (PACKSADDLE SCHIST)	Metasedimentary Rocks
					ROUGH RIDGE FM. (PACKSADDLE SCHIST)	
					SANDY FM. (PACKSADDLE SCHIST)	
					HONEY FM. (PACKSADDLE SCHIST)	
					VALLEY SPRING GNEISS	
					LOST CREEK GNEISS	
					COAL CREEK SERPENTINE BIG BRANCH GNEISS RED MOUNTAIN GNEISS	Meta-igneous Rocks
					Quartz porphyry (LLANITE) SIXMILE GRANITE OATMAN CREEK GRANITE TOWN MOUNTAIN GRANITE	Igneous Rocks

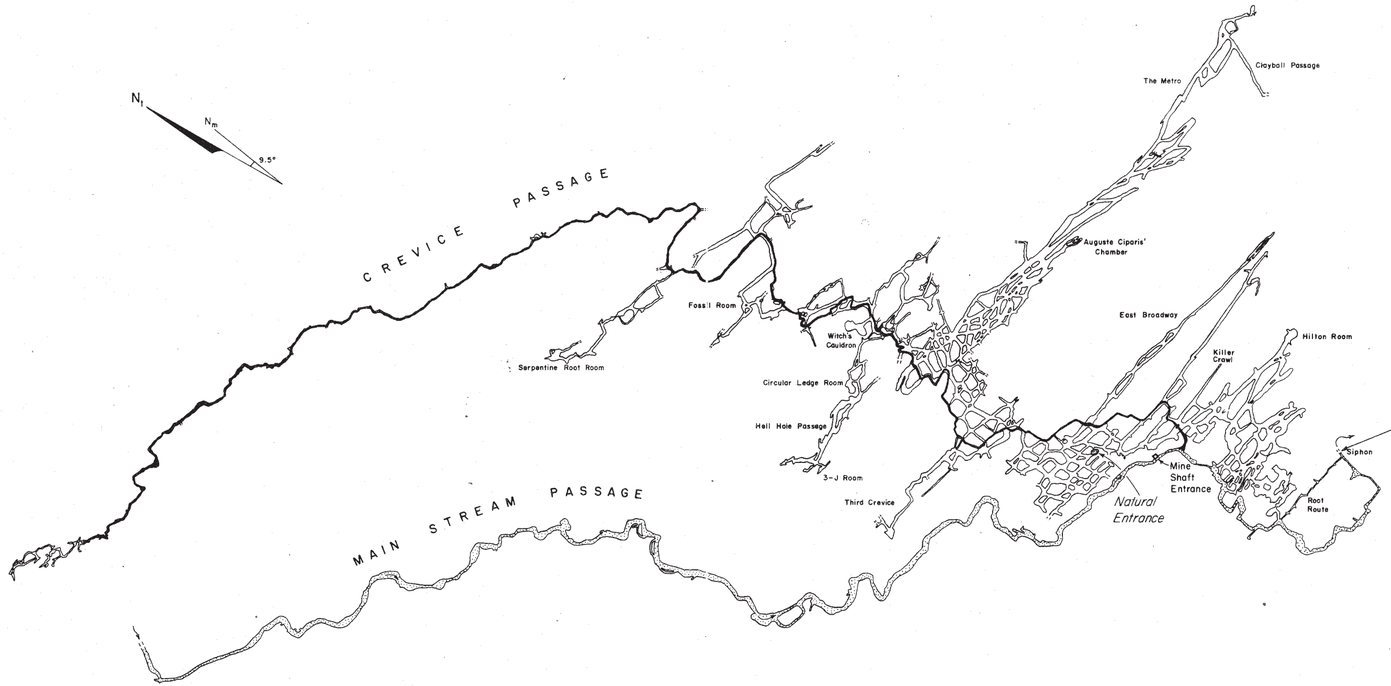
E. H. Kastning 1983

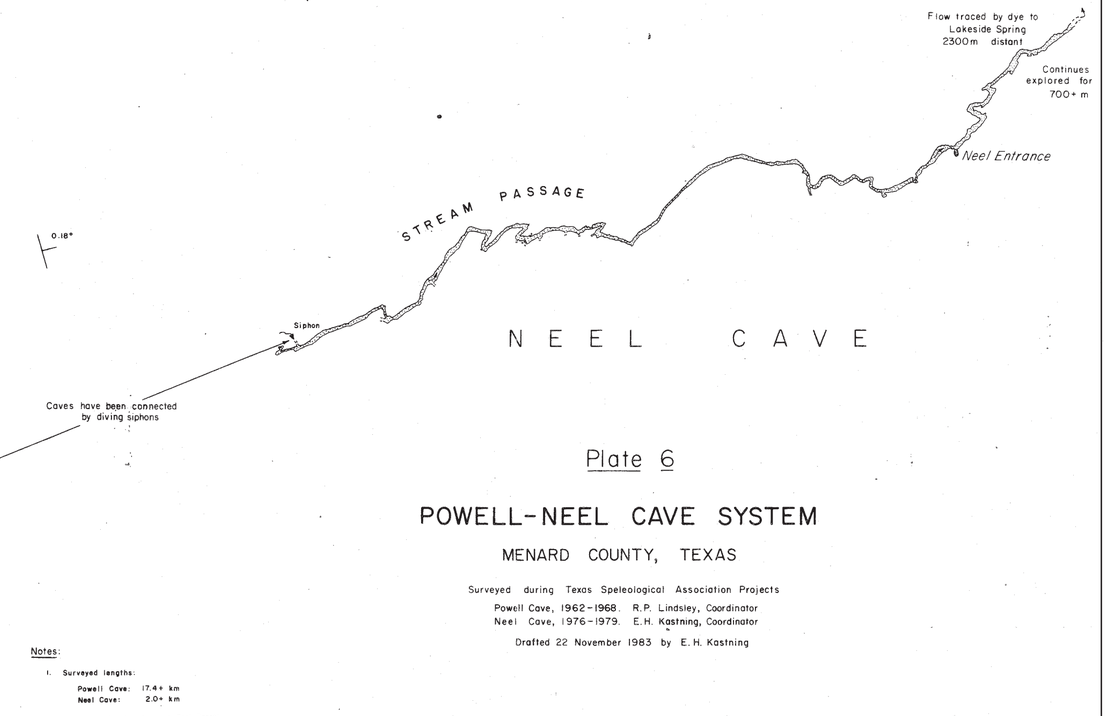
Plate 5

STRATIGRAPHIC CORRELATION CHART
CRETACEOUS ROCKS
EDWARDS PLATEAU, CENTRAL TEXAS



E. H Kastning 1983





Notes:

1. Surveyed lengths:
   Powell Cave: 17.4+ km
   Neel Cave: 2.0+ km
2. Caves are developed in the Fort Terrett Formation of the Edwards Group (Lower Cretaceous).
3. This map is simplified from manuscript maps drawn at scale of 1:600 (1 inch = 50 feet).

Plate 6

POWELL-NEEL CAVE SYSTEM

MENARD COUNTY, TEXAS

Surveyed during Texas Speleological Association Projects

Powell Cave, 1962-1968. R.P. Lindsley, Coordinator
Neel Cave, 1976-1979. E.H. Kastning, Coordinator

Drafted 22 November 1983 by E.H. Kastning

50 0 50 100 150 200 250 300 350 meters

200 0 200 400 600 800 1000 1200 feet