KINEMATICS OF THE FRONTAL HIMALAYAN THRUST BELT, PAKISTAN, AND THE EXTERNAL WESTERN ALPS, FRANCE.

by

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ABSTRACT.

On a large scale, thrust–and–fold belts are characterised by **arcuate zones** in which the direction of overthrusting (or tectonic transport direction) is divergent, e.g. the Western Alps of SE. France, and **syntaxial zones** where transport directions are convergent, e.g. the Hazara–Kashmir Syntaxis of N. Pakistan. Such **transport direction variations** result from the original shape of the colliding continental margins, the relative plate motion vector and gravitational body forces. On a smaller scale, transport direction variations in the external zones of thrust–and–fold belts are dependent on the inherited structural and stratigraphic framework of the under–riding continental margin.

The tectonic transport direction may vary with time during orogeny and spatially within the orogen. Two zones of non–colinear thrust interference have been studied. One, in the Pelvoux basement massif of the external French Alps, results from a temporal variation in transport direction involving the superposition of two successive deformation phases during the propagation of a single thrust wedge. The other, in the Hazara–Kashmir Syntaxis of N.Pakistan, results from spatial variation in transport direction involving the interference of two concurrent, but convergent, thrust systems.

Two phases of Alpine deformation can be distinguished in the environs of the Pelvoux massif due to the presence of an unconformable Nummulites–bearing flysch series. The **Pre–Nummulitic phase** was generally W– to WSW–vergent and involved the ductile folding of basement and cover imbrication. Major uplift that exposed the basement is associated with the inversion of Mesozoic half–graben and the reactivation of their bounding faults. The **Post–Nummulitic phase** is characterised by breaching thrust geometries in both the basement and cover series. The earlier uplift of basement meant that, during this phase, the Pelvoux massif acted as a resistant block. Further tightening of structures within the massif occurred, but local stress anomalies caused variations in transport direction between WSW and NW. In the flanking cover sequences to the north and south of the massif transport directions have been deflected

towards the WNW and WSW respectively. Both phases were strongly influenced by inherited Mesozoic extensional basement–cover geometries.

Himalayan deformation on the western limb of the Hazara–Kashmir Syntaxis is weakly influenced by offsets in the basement. Instead, the gently–flexed, basement–cover interface has been exploited as a décollement horizon. The SSE–vergent deformation in the external zones of the Himalayan thrust wedge is therefore restricted to the sedimentary cover sequence. Four deformation phases, that accommodate over 200km shortening, are recognised. Each of these phases is characterised by a different structural style.

Major lateral variations in structural style across the Potwar Plateau also occur. These have been attributed mainly to the erosion of the Mesozoic shelf sequence beneath the base Palaeocene unconformity, but basement dip variations may also be a factor. Uplift associated with the passage of India over the Reunion hotspot is proposed as an origin for the base Palaeocene unconformity.

SW–vergent thrusting of the Kashmiri thrust wedge has been superimposed upon this SSE–vergent deformation. This has resulted in interference, rotation and additional uplift of the region which has been eroded to reveal an arcuate outcrop pattern. The syntaxis is seen as a product of the original shape of India at collision. Collision occurred first at the northwestern tip of India, and Himalayan thrusting migrated laterally from this point. The Main Himalayan arc has always been pinned by the Pakistani Himalaya at the Northwest Syntaxis. This pinning has produced convergent transport directions, the active expression of which is the Hazara–Kashmir Syntaxis.

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TABLE OF CONTENTS.

ABSTRACT 2

ACKNOWLEDGEMENTS 4

CONTENTS 7

LIST OF FIGURES 10

LIST OF PLATES 17

CHAPTER 1. INTRODUCTION. 19

1.1. AIMS AND OBJECTIVES. 19

1.2. METHODOLOGY. 19
1.2.1. Remote Sensing Data. 19
1.2.2. Geophysical Data. 20
1.2.3. Well Data. 20
1.2.4. Structural Field Mapping. 22
1.2.5. Logistics. 22

1.3. OUTLINE OF THESIS. 23

CHAPTER 2. THRUST BELT THEORY AND TERMINOLOGY. 26

2.1. THRUST GEOMETRIES. 27

2.2. THRUST WEDGE PROPAGATION THEORY. 35

2.3. THRUST INTERFERENCE. 38

2.4. RHEOLOGY AND STRUCTURAL STYLE. 39

2.5. KINEMATIC INDICATORS. 41

2.6. STRUCTURAL CROSS-SECTION CONSTRUCTION. 44

CHAPTER 3. AN INTRODUCTION TO THE WESTERN ALPS. 47

3.1. ALPINE RESEARCH AND THE APPLICATION OF PLATE TECTONICS. 47

3.2. EVOLUTION OF THE SOUTHERN EUROPEAN MARGIN OF TETHYS. 50

3.3. STRATIGRAPHY OF THE DAUPHINOIS ZONE. 56

3.4. STRUCTURAL FRAMEWORK OF THE DAUPHINOIS ZONE. 66

3.5. PHASES OF ALPINE DEFORMATION. 68
3.5.1. The Pre-Senonian Phase. 68
3.5.2. The Pyrenean-Provencal Phase. 69
3.5.3. The Pre-Nummulitic Phase. 69
3.5.4. The Post-Nummulitic Phase. 74

CHAPTER 4. FIELD DATA FROM THE WESTERN FRENCH ALPS. 79

4.1. FIELD DATA FROM SOUTHERN PELVOUX. 81
4.1.1. St.Bonnet/Les Infournas Hauts. 87
4.1.2. La Muande. 90
4.1.3. Tourond and Cedera. 95

4.2. FIELD DATA FROM SOUTHEASTERN PELVOUX. 109
4.2.1. Méollion/Rougnoux. 112
4.2.2. Selle. 125
4.2.3. Fournel. 131
4.2.4. Dourmillouse. 133

4.3. FIELD DATA FROM EASTERN PELVOUX. 140
4.3.1. Chambran/l'Yret. 140
4.3.2. La Blanche. 144
4.3.3. Ailefroide. 148

4.4. FIELD DATA FROM THE ARVINCHE REGION, NORTH OF PELVOUX. 150
4.4.1. The Arvan Valley. 154
4.4.2. The Aiguilles d'Arves–Pointe des Chaudannes Ridge. 165
4.4.3. Summary of the Arvinche deformation History. 168

CHAPTER 5. INTERFERENCE OF ALPINE PHASES IN PELVOUX, S.E.FRANCE. 170

5.1. MESOZOIC EXTENSION. 170

5.2. PRE–NUMMULITIC ALPINE DEFORMATION. 175
5.2.1. Pre–Nummulitic Deformation in the Arvinche. 175
5.2.2. Pre–Nummulitic Deformation in the Pelvoux massif. 178

5.3. POST–NUMMULITIC ALPINE DEFORMATION. 181

CHAPTER 6. AN INTRODUCTION TO THE HIMALAYAS. 184

6.1. THE PLATE TECTONIC SCENARIO. 184

6.2. ZONATION OF THE HIMALAYAS. 186

6.3. THE INTERNAL ZONES OF THE HIMALAYAS. 189

6.4. THE EXTERNAL ZONES OF THE HIMALAYAS. 190
6.4.1. The Galiat. 191
6.4.2. The North Potwar Deformation Zone. 192
6.4.3. The Southeastern Potwar. 193
6.4.4. THe Hazara–Kashmir Syntaxis. 194

6.5. STRATIGRAPHIC EVOLUTION OF THE EXTERNAL ZONES. 195
6.5.1. The Wilson Cycle : Gondwana Break–up to Collision. 196
6.5.2. The Base Palaeocene Unconformity. 199
6.5.3. Continent–Continent Collision. 202
6.5.4. Post–Collisional Convergence. 205

6.6. THE STRATIGRAPHY OF THE GALIAT AND POTWAR REGIONS OF PAKISTAN. 206
6.6.1. The Eocambrian. 210
6.6.2. The Lower Palaeozoic. 217
6.6.3. The Permian–Cretaceous. 222
6.6.4. The Lower Tertiary. 231
6.6.5. The Upper Tertiary. 237

CHAPTER 7. FIELD DATA FROM THE PAKISTANI HIMALAYAS. 246

7.1. STRUCTURAL DATA FROM THE GALIAT. 250
7.1.1. Murree. 252
7.1.2. Kuldanna–Khaira Gali : the Nummulitic Zone. 266
7.1.3. Khaira Gali–Kuza Gali : the environs of Changla Gali. 272
7.1.4. Kuza Gali–Dunga Gali : the Ayubia National Park. 277
7.1.5. Dunga Gali–Bagnotar : the Slate Zone. 283
7.1.6. Summary of the Galiat Region. 290

7.2. STRUCTURAL DATA FROM THE N.P.D.Z. 292
7.2.1. The Bharakao–Thaliala Section. 294
7.2.2. The Danah–Lehtrar Section. 299
7.2.3. The Kashmir Point–Nara Secton. 302
7.2.4. Transport Direction Indicators. 305
7.2.5. Summary of the Eastern NPDZ Region. 307

7.3. STRUCTURAL DATA FROM THE SOUTHEASTERN POTWAR. 309
7.3.1. Riwat. 311
7.3.2. Buttar. 314
7.3.3. Ghungrila–Kallar. 315
7.3.4. Qazian. 318
7.3.5. Domeli. 323
7.3.6. Jogi Tilla Thrust/Mahesian. 325
7.3.7. Rohtas. 330
7.3.8. Pabbi. 332
7.3.9. Summary of the SE. Potwar. 333

CHAPTER 8. AN EVOLUTIONARY MODEL FOR THE WESTERN LIMB OF THE HAZARA–KASHMIR SYNTAXIS. 334

8.1. FACTORS CONTROLLING THE GEOMETRY OF HIMALAYAN GEOMETRY. 334

8.2. THE FORMATION OF THE HAZARA–KASHMIR SYNTAXIS. 337

8.3. DEFORMATION HISTORY OF THE WESTERN LIMB OF THE HKS. 341
8.3.1. Imbrication of the Hangingwall to the Hazara Detachment. 345
8.3.2. Translation on the Hazara Detachment (MBT). 347
8.3.3. Duplex Formation in the Footwall to the MBT. 351
8.3.4. Displacement on the Salt Range Detachment (SRT). 355

8.4. VARIATIONS IN STRUCTURAL STYLE ACROSS THE POTWAR PLATEAU. 359

8.5. THRUST WEDGE INTERFERENCE AND THE FORMATION OF THE HKS. 364

CHAPTER 9. CONCLUSIONS. 370

9.1. THE EXTERNAL FRENCH ALPS. 370
9.1.1. The Pre-Nummulitic Phase of the Arvinche. 370
9.1.2. The Pre-Nummulitic Phase of Pelvoux. 371
9.1.3. The Post-Nummulitic Phase. 371
9.2. THE EXTERNAL HIMALAYAS OF PAKISTAN. 373

BIBLIOGRAPHY. 376

LIST OF FIGURES.

CHAPTER 1.

Fig.1.1 Map showing Location of Landsat MSS and Aerial Photograph coverage, N. Pakistan. 21

Fig.1.2 Map showing the Topographic Map Database available, N.Pakistan. 24

CHAPTER 2.

Fig.2.1 Block diagram illustrating the basic elements of a thrust system. 28

Fig.2.2a. Diagram illustrating thrust footwall geometries. 29

Fig.2.2b. Diagram illustrating thrust hangingwall culminations. 29

Fig.2.3. Fault-bend folding model. 30

Fig.2.4. Diagram illustrating breaching thrust geometry. 32

Fig.2.5. Diagram illustrating dupliex formation. 32

Fig.2.6. Styles of fault tip. 33

Fig.2.7. Out-of-syncline thrusting. 33

Fig.2.8a. Pop-up formation. 34

Fig.2.8b. Triangle zone formation. 34

Fig.2.9. Shear sense indicators. 43

CHAPTER 3.

Fig.3.1 The Relationship between Europe/Africa Plate Motions and Alpine Events. 49

Fig.3.2 The Disparity between Alpine Kinematic Data and Plate Motion Vectors. 51

Fig.3.3 Map showing the Zonation of the Western Alps. 52

Fig.3.4 Cartoon illustrating the Palaeogeographic Domains of the Western Alps. 54

Fig.3.5 Subsidence Curves for the Palaeogeographic Domains of the European Margin, SE.France. 55

Fig.3.6 Bathymetric Evolution of the Tethyan Margin during the Jurassic. 57

Fig.3.6a Basement Framework during Tethyan rifting in the Western Alps. 58

Fig.3.7 Timing of Rifting in the Pelvoux Massif. 59

Fig.3.8 Sketch-section illustrating the Basin Structure of the Western Alps. 62

Fig.3.9 Halokinesis during Syn-rift Sedimentation on the Tethyan Margin of the W.Alps. 64

Fig.3.10 Superposition of Alpine Phases in Pelvoux. 71

Fig.3.11 Fold/Thrust Relationships. 72

Fig.3.12 Inversion of Mesozoic Half-graben in N.Pelvoux. 73

Fig.3.13 Tilting of the Nummulitic Unconformity in S.Pelvoux: Evidence of Pre-Nummulitic uplift. 76

Fig.3.14 Thrust Kinematic Data from the Western Alps. 77

Fig.3.15 Possible Mechanisms of Thrust Transport Direction Variation in the W.Alps. 78

CHAPTER 4.

Fig.4.1 Map of the Pelvoux Basement Massif. 80

Fig.4.2 Map of the Champsaur Region, Southern Pelvoux. 82

Fig.4.3 Map of Southern Pelvoux. 84

Fig.4.4 Partialy Restored Cross-Sections across Southern Pelvoux, along Line A-A'. 85

Fig.4.5 Map of the St.Bonnet locality, Southern Pelvoux. 88

Fig.4.6 Map of the Muande locality, Southern Pelvoux. 91

Fig.4.7 Schematic Strike Section through the Muande locality. 94

Fig.4.8 Map of the Tourond and Cedera areas, Southern Pelvoux. 97

Fig.4.10a Section across the Tourond-Cedera Structure. 98

Fig.4.10b Section across the Tourond-Cedera Structure. 99

Fig.4.10c Section across the Tourond Structure. 100

Fig.4.10d Section across the Tourond Structure. 101

Fig.4.10e Section across the Tourond Structure. 102

Fig.4.11 Cartoons showing the Geometry of the Tourond Thrust and its Footwall Imbricates. 105

Fig.4.13 Map of Southeastern Pelvoux. 110

Fig.4.14	Map of the Méollion-Rougnoux locality, southeastern Pelvoux.	113
Fig.4.15	Section across the Méollion-Rougnoux area.	114
Fig.4.16	Section across the Méollion-Rougnoux area.	114a
Fig.4.17	Field Sketch of the Eocene Flysch exposure at Crete du Martinet, Rougnoux.	118
Fig.4.18	Cartoon illustrating the Deformation Style within the Flysch sequence.	119
Fig.4.19	Cartoon illustrating the proposed Tilted Fault Block Geometry prior to Post-Nummulitic Thrusting.	122
Fig.4.20	Sketch Section through Lateral HW ramp of Sirac Thrust.	123
Fig.4.21	Map of the Selle locality, SE. Pelvoux.	126
Fig.4.22	Section through the Selle structure.	127
Fig.4.23	Map of the Fournel and Dourmillouse localities, SE. Pelvoux.	132
Fig.4.24	Two Sections across the Dourmillouse-Méollion structure.	134
Fig.4.25	Section across the Selle, Fournel and Dourmillouse structures.	137
Fig.4.26	Cartoon illustrating the possible Pre-Nummulitic structure between Selle and Dourmillouse.	138
Fig.4.27	Map of the La Blanche, Chambran-l'Yret and Ailefroide localities, E. Pelvoux.	141
Fig.4.28	Strike Section the FPT Zone at Chambran, E. Pelvoux.	142
Fig.4.29	Section through La Blanche, E. Pelvoux.	145
Fig.4.30	Cartoons illustrating the Relationship between the Eocene Flysch Series and the FPT.	147
Fig.4.31	Section through the Ailefroide locality, E. Pelvoux.	149
Fig.4.32	Map of the Arvinche region, North of Pelvoux.	151
Fig.4.33	Stratigraphic Columns from the Ultra-Dauphinois Thrust Sheet.	153
Fig.4.34	Map of the Arvinche Fold System.	155
Fig.4.35	Section across the Ultra-Dauphinois Zone.	157

Fig.4.36 Cartoon illustrating a possible Origin of the Fanning of the Arvinche Folds. 158

Fig.4.37 Dip Section through the Ultra-Dauphinois Thrust Sheet. 160

Fig.4.38 Cut-away Block Diagram illustrating the Structure of the Arvinche region. 161

Fig.4.39 Section through the Arvan Gorge. 163

Fig.4.40 Section along the Aiguilles d'Arves-Grand Chible ridge. 164

Fig.4.41 Field Sketch of the Eocene Flysch outcrops, East of Roche de Bonhomme. 167

CHAPTER 5.

Fig.5.1 Block diagram illustrating the basement geometry of the Cedera half-graben after Liassic rifting. 172

Fig.5.2 Sections illustrating the phases of structural evolution in S.Pelvoux. 173

Fig.5.3 Map illustrating the deformation phases in the Pelvoux massif. 176

CHAPTER 6.

Fig.6.1 The Zonation of the Himalayas. 187

Fig.6.2 Sketch Section showing the Zonation of the Pakistan Himalayas. 188

Fig.6.3 Gondwana Break-up and India's Passive Margin Development. 197

Fig.6.4 Precambrian Lineament Trends of Gondwana. 198

Fig.6.5 The Drift of India and the Reunion Hotspot : Control on the Basal Palaeocene Unconformity. 201

Fig.6.6 Motion of India since 70Ma, showing Deceleration and continued Convergence since Collision. 204

Fig.6.7 Schematic N-S Stratigraphic Correlation Chart at 73° 30'E Longitude. 207

Fig.6.8 Diagrammatic E-W Stratigraphic Correlation Chart across the Salt Ranges. 208

Fig.6.9 Gondwanaland Reconstruction showing the Distribution of Eocambrian Evaporites in Pakistan and Arabia. 216

Fig.6.10 Distribution of Late Precambrian to Cambrian Sediments in Hazara and the Potwar. 223

Fig.6.11 Basal Palaeocene Subcrop Map. 229

Fig.6.12 Chronostratigraphic Chart across the Potwar at Latitude 33°N. 235

Fig.6.13 Bouguer Gravity Map of the Northern Punjab. 241

Fig.6.14 Stratigraphic Correlation in N. Pakistan. 219

CHAPTER 7.

Fig.7.1 Locality Map of the Hazara-Kashmir Syntaxis. 247

Fig.7.2 Topography, Drainage and Communications in the Galiat. 251

Fig.7.3 Geological Map of Murree showing the geometry of the Murree Thrust. 253

Fig.7.4a Section across the Murree Thrust at Bansara Gali/Pindi Point. 254

Fig.7.4b Section across the Murree Thrust at Chitta Mor Gali/Murree. 255

Fig.7.4c Section across the Murree Thrust at Kuldanna/ Kashmir Point. 256

Fig.7.5 Simplified Structural Re-interpretation of Latif's (1970) Map of S.E. Hazara. 258

Fig.7.6 Simplified Structural Re-interpretation of Latif's (1970) Sections across S.E. Hazara. 259

Fig.7.7 Map and Section of the MBT Zone at Tret, on the Pindi-Murree Road. 263

Fig.7.8 Cartoon illustrating the Evolution of the MBT Zone at Tret. 264

Fig.7.8b Structural Data from the Hazara Detachment (MBT) at Kohala. 265

Fig.7.9 Map of the Nummulitic Zone between Khaira Gali and Kuldanna. 267

Fig.7.10 Section through the Nummulitic Zone between Khaira Gali and Kuldanna. 268

Fig.7.11 Map of the area between Kuza Gali and Khaira Gali. 273

Fig.7.12a Section through Changla Gali, west of the main road. 274

Fig.7.12b Road Section from Kuza Gali to Khaira Gali. 275

Fig.7.13 Map of Dunga Gali and Ayubia. 278

Fig.7.14a Road Section from Kundla to Kuza Gali 279

Fig.7.14b Section through Ayubia National Park from Dunga Gali to Khanspur Cantonment. 280

Fig.7.15 Map of the Slate Zone 284

Fig.7.16a Section A-A' across the Slate Zone from Bagnotar to Kalabagh Cantonment. 285

Fig.7.16b Section B-B' from Nathia Gali to Dunga Gali. 286

Fig.7.17a Parasitic Backfolds verge towards inferred major axis, near Ghia. 289

Fig.7.17b Sketch of Backfolds between Toydabad and Kundla. 289

Fig.7.18 Map of the uplifted Eastern NPDZ, South of Murree. 293

Fig.7.19a Section across the eastern NPDZ between Bharakao and Kirpa. 295

Fig.7.19b Section across the eastern NPDZ between Danah and Lehtrar. 300

Fig.7.19c Section across the eastern NPDZ between Kashmir Point and Nara. 303

Fig.7.20 Cartoon and Detailed Field Sketch illustrating the structure at Thanda Pani. 296

Fig.7.21 WSW-ENE trending Sketch Section across the Eastern NPDZ. 304

Fig.7.22 Lineation Data Plot for the Eastern NPDZ. 306

Fig.7.23 Surface Data from the Eastern Potwar Plateau 310

Fig.7.24 Interpretations of Seismic Time Sections 1 & 2, incorporating surface data. 312

Fig.7.25 Evolution of the Riwat Fault. 313

Fig.7.26 TWT Contour Map of the Base Molasse, SE Potwar. 316

Fig.7.27 Fault Cut-Off Map, SE Potwar, at the Base Molasse Contact. 317

Fig.7.28 Interpretations of Seismic Time Sections 3 & 4, incorporating surface data. 319

Fig.7.29 Interpretations of Seismic Time Sections 5 & 6, incorporating surface data. 320

Fig.7.30 Interpretations of Seismic Time Sections 7 & 8, incorporating surface data. 321

Fig.7.31 Along Strike Variations in Surface Structure at Qazian. 322

Fig.7.32 Variation of the Domeli Structure along strike. 325

Fig.7.33 Variation of the Jogi Tilla-Mahesian Structure along strike. 327

Fig.7.34 Interpretation of Seismic Time Section "S". 328

Fig.7.35 Interpretation of Seismic Time Section 9, incorporating surface data. 329

CHAPTER 8.

Fig.8.1 Map showing the effects of Himalayan Orogeny in Southern Asia. 335

Fig.8.2 Distribution of Palaeomanetic Vector Azimuths around the Hazara-Kashmir Syntaxis. 340

Fig.8.3 Simplified Structural Map of the Eastern Potwar and Southern Hazara. 343

Fig.8.4 Structural Cross-Section across the Hil Ranges and the Eastern Potwar. 344

Fig.8.5 Deformed and Restored Cross-Sections across the Galiat from Bagnotar to Murree. 348

Fig.8.6 Contrasting Models of MBT Emplacement. 350

Fig.8.7 Cartoon illustrating the "Steady-State" Propagation of the MBT Wedge. 352

Fig.8.8 Cartoon illustrating the Basic Structure of the NPDZ. 354

Fig.8.9 Contrasting Wedge Character and the Effect of Salt. 356

Fig.8.10 Predicted Trail of the Reunion Hotspot. 363

Fig.8.11 Thrust Interference and Timing in the Hazara-Kashmir Syntaxis. 366

Fig.8.12 Structure of the Hazara-Kashmir Syntaxis. 367

LIST OF PLATES

Plate 1.	The Outcrop of the Vieux Chaillol Thrust at Muande.	92
Plate 2a.	View of Aiguille de Cedera from the south, showing the Nummulitic unconformity.	116
Plate 2b.	View of sheared Triassic spilite overlain by Eocene conglomerate, at Col du Cheval de Bois.	116
Plate 3.	View of the Sirac and Méollion Thrust sheets at Col de Méollion (strike-section).	117
Plate 4.	View of the Rouite Fault from the west.	121
Plate 5.	View of the Col du Cheval de Bois from the east, showing Sirac Thrust sheet (strike-section).	124
Plate 6.	View of the northeastern part of the Selle syncline, showing evidence of Pre- and Post-Nummulitic deformation.	129
Plate 7.	View of Post-Nummulitic deformation at Col de l'Aup Martin.	130
Plate 8a.	Transport Direction indicators at Dourmillouse.	139
Plate 8b.	Principal Stress Direction indicators at Dourmilouse.	139
Plate 9.	The Nummulitic unconformity at La Blanche.	146
Plate 10a.	Hazara Group outcrop showing bedding-parallel foliation.	211
Plate 10b.	Hazara Group outcrop showing sole casts.	211
Plate 11a.	Hazara Group outcrop showing bedding-perpendicular conjugate jointing.	212
Plate 11b.	Hazara Group outcrop showing kinking.	212
Plate 12.	Deformation within the Miranjani Lst Member of the Hazara Group.	213
Plate 13.	Mylonitic fabric at the Hazara Detachment (MBT) just S. of Muzaffarabad.	214
Plate 14.	The Tanakki boulder bed unconformably overlying Hazara Group greywackes.	218
Plate 15a.	The basal Permian unconformity near Khewra.	221
Plate 15b.	The Cambrian Jhelum Group sequence near Khewra.	221
Plate 16.	The Jurassic Samana Suk Formation in Hazara.	226

Plate 17. Fold Hinge Deformation in Samana Suk Formation, NW. of Bagnotar. 227

Plate 18. Basal Kawagarh Formation, near Kundla. 228

Plate 19a. Cretaceous-Tertiary boundary, with Hangu Fm. laterite, near Changla Gali. 230

Plate 19b. Kawagarh Formation, near Changla Gali (c.f. Plate 18). 230

Plate 20a. Palaeocene Patala Formation shales. 233

Plate 20b. Eocene Margalla Hil Formation nodular, nummulitic limestone. 233

Plate 21a. Eocene Chora Gali Formation platy, nodular nummulitic limestone. 234

Plate 21b. Eocene Kuldanna Formation evaporitic, red marls and limestones. 234

Plate 22a. Miocene Murree Formation fluvial molasse. 238

Plate 22b. Fatehjang Member detrital Nummulites bed (marking the transition from shelf to foreland basin sedimentation). 238

Plate 23a. Siwalik Group sediments at Chambal Ridge, eastern Salt Range. 243

Plate 23b. Soan Formation conglomerates, eastern Potwar. 243

Plate 24a. Shear zone within the Chora Gali Formation. 269

Plate 24b. Flexural slip and tangential-longitudinal strain folding in the interbedded shales and limestones of the Margalla Hil Formation. 269

Plate 25. Backsteepened SE-vergent shears in the Galiat. 257

Plate 26. The Salt Range Thrust near Jalalpur, eastern Salt Range. 248

Plate 27. Monoclinally-folded Siwalik sediments in front of the Salt Range Thrust, eastern Salt Range. 249

CHAPTER 1.

INTRODUCTION.

1.1 AIMS AND OBJECTIVES.

The main objectives of this project were to examine the structural style of deformation and to chart the structural evolution of zones of tectonic transport direction variation within thrust–and–fold belts. A further aim was to constrain the role of inherited stratigraphic variations upon structural style. Both the Alpine and Himalayan field areas provide excellent examples of such variations. A particular aim of the Himalayan research has been to develop a model of the structural evolution of the Hazara–Kashmir Syntaxis leading to more general concepts of syntaxis development in orogenic belts.

Hence, the thesis aims to :

1) constrain the role of transport direction variation in controlling structural style,

2) emphasise the role of stratigraphic variation in controlling structural style,

3) apply the techniques of thrust tectonics to a region of marked non–plane strain and

4) present a model for the structural evolution of the western limb of the Hazara–Kashmir Syntaxis, N. Pakistan.

1.2 METHODOLOGY.

1.2.1 Remote Sensing Data.

Remote sensing data have proved very useful in parts of the Pakistani field area. The large scale structures of the external Himalayas, particularly in the Potwar region, have been successfully imaged on Landsat MSS data. Aerial photos, held at the Geological Department of the University of the Punjab, Lahore, proved helpful in establishing the continuity of structures already observed in the field. Their applicability to the reconnaisance mapping of the Hill Ranges

is minimal due to the complexity of the geological structure, however, in the less intensely deformed Potwar region the photos permit structures to be readily imaged (Fig.1.1).

In the French Alps, remote sensing techniques have not been used due to the extent of previous detailed geological mapping and the scale of the field relationships studied.

1.2.2 Geophysical Data.

Seismic reflection data covering the Potwar Plateau region have been acquired by numerous companies since the mid-1970's. The generosity of certain companies made access to this data possible. This proved invaluable to the refinement of the structural cross-sections across the eastern Potwar Plateau area. The seismic coverage does not extend northwards into the Hill Ranges, however, and is of poor quality in the northern Potwar. The data for the southern Potwar are of good quality.

Towards the north only published magnetic and gravity survey data and natural seismicity data place geophysical constraints upon the deep structure.

In the western Alps a deep seismic traverse has been acquired by the ECORS group and magnetic and gravity data are also available. These data permit the gross basement structure and deformation style of the external French Alps to be constrained.

1.2.3 Well Data.

Access to commercial well data was essential to constrain sections through the Potwar Plateau of Pakistan, particularly with regard to mapping the unconformities and the sub-molasse strata. The rapid stratigraphic variation and extent of tectonic telescoping to the north of the Potwar Plateau makes extrapolation of these data inaccurate. Although helpful in the estimation of molasse thickness, the lesson of field studies is that well data should be treated with caution where the dating of the molasse is concerned. The transgressive nature of the molasse deposits makes lithostratigraphy of limited value and can lead to misinterpretation of well data.

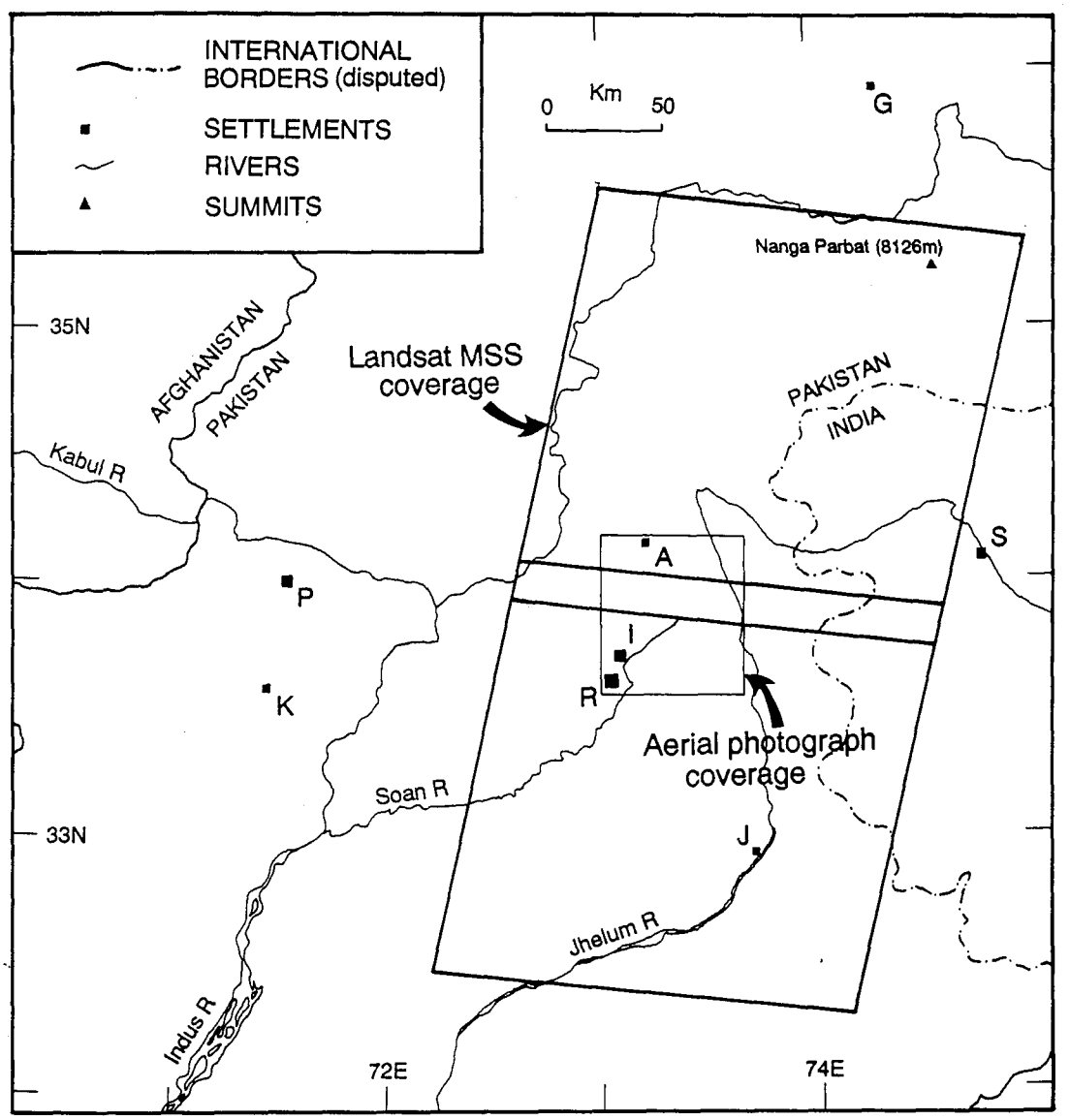


Fig.1.1. LANDSAT MSS AND AERIAL PHOTOGRAPH DATA.

G, Gilgit; P, Peshawar; K, Kohat; A, Abbottabad; I, Islamabad; R, Rawalpindi; J, Jhelum; S, Srinigar

1.2.4 Structural Field Mapping.

The basis of this project has been data acquired during structural field mapping in the French Alps and the Pakistani Himalayas. In total, three visits have been made to each field area as follows :

France :	June–July, 1987	7 weeks
	July–August, 1988	7 weeks
	June–July, 1989	3 weeks
Pakistan :	October–December, 1987	8 weeks
	October, 1988–January, 1989	19 weeks
	March–May 1990	12 weeks

The topographic relief, in places in excess of 2km, leads to excellent exposure of the geological structure.

In both field areas the nature of the fieldwork was to map in detail the structural relationships exposed. The scale of the field areas, especially in Pakistan, has prohibited the production of comprehensive geological maps.

1.2.5 Logistics.

The logistic problems posed by working in the French Alps and the Pakistani Himalayas are incomparable. In France, the Alps are covered by an excellent series of maps at 1 : 25 000 scale produced by the Institut Geographique National and access to the sparsely populated or uninhabited hills is unrestricted.

In Pakistan, maps are available from the Survey of Pakistan, Rawalpindi, at 1 : 50 000 and 1 : 250 000 scales but for most of the area studied can not be purchased for security

reasons (Fig.1.2). Official sanction to carry out fieldwork in Pakistan takes the form of "No-Objection Certificates", which were obtained from the Ministry of Education (1987, 1988/89) and the Ministry of Petroleum (1990).

Vehicular access to the Hill Ranges and the Potwar Plateau is slow due to a poor communications infrastructure. However, 4WD jeeps permit access to most areas, although in the rainy season minor roads on the Potwar Plateau become impassable and in the winter the snowline encroaches upon the Hill Ranges.

The Hill Ranges and the Potwar Plateau are heavily populated. Far from causing problems this ensures that, even at 2500m, a network of roads and paths exists and that accommodation and refreshment stops are readily available. Accommodation in the Pakistan study area comprises hotels (this is one of Pakistan's prime tourist centres) and Government Resthouses, particularly those of the Forestry Department which are superbly sited. Camping is out of the question.

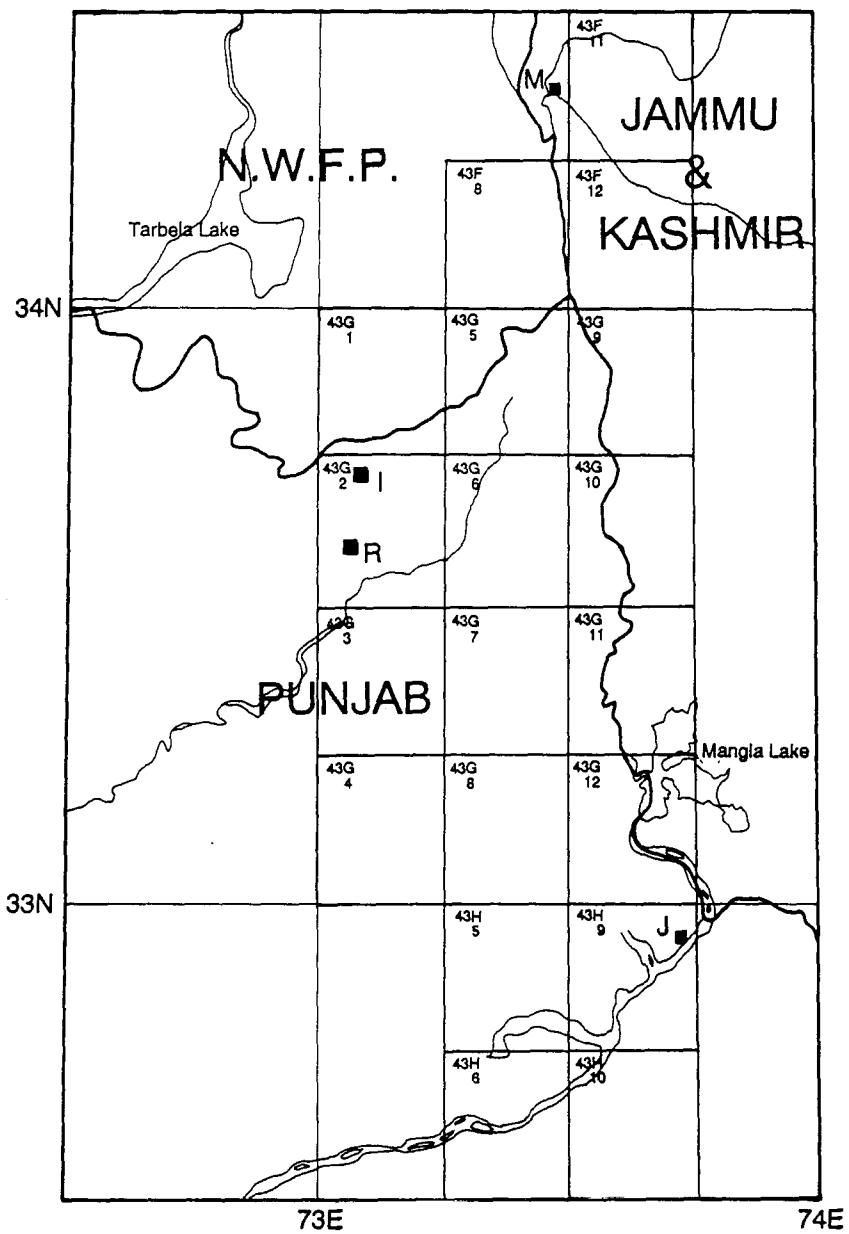
1.3. THESIS OUTLINE.

After an introduction to the pertinent theory (Chapter 2), the research is presented in two halves.

In the first half the style of thrust interference between successive phases of Alpine deformation in the Pelvoux massif, France, is presented. Previous work in the region is described (Chapter 3), then the new data are presented (Chapter 4) and briefly discussed (Chapter 5).

In the second half, the evolution of the western limb of the Hazara-Kashmir Syntaxis, Pakistan, and the interference of two concurrent, but convergent, thrust systems is presented. Again, the existing knowledge of the region is initially described (Chapter 6), then the

Fig.1.2. TOPOGRAPHIC MAP DATABASE AT 1:50 000 FOR THE WESTERN LIMB OF THE HAZARA-KASHMIR SYNTAXIS.



M, Muzaffarabad; I, Islamabad; R, Rawalpindi; J, Jhelum.

new data are presented (Chapter 7) and interpreted (Chapter 8). Finally, the conclusions of the work are stated (Chapter 9).

CHAPTER 2.

THRUST BELT THEORY AND TERMINOLOGY.

The aim of this chapter is to introduce pertinent geological theory and terminology relating to the development of fold–and–thrust belts, prior to presenting the data from the Alpine and Himalayan belts.

Initially a brief summary of thrust systems (Fig.2.1) and thrust geometry is presented, providing the essential foundation upon which the project has been built. This is followed by a review of wedge propagation theory, in which the progressive nature of thrust belt deformation is emphasised, leading to the consideration of thrust reactivation, out–of–sequence thrusting and thrust interference patterns.

The influence of rheology upon deformation style is introduced, particularly with respect to the presence of salt in thrust belts. Regional stratigraphic variations caused by the action of unconformities or pre–existing structure lead to variations in the structural style of the thrust belt.

During the course of the project the determination of tectonic transport directions in the field has been vital. The interpretation of kinematic indicators is discussed. Structural cross–sections are traditionally drawn parallel to the tectonic transport direction because of the plane strain assumption inherent in the section–balancing technique. A description of the various methods of section balancing is given and an assessment of their applicability made, particularly where non–plane strain conditions prevail. Restored sections must reflect the pre–orogenic structure.

In summary, a review of the following topics is given :

1) the basics of thrust tectonics,

2) thrust wedge propagation,

3) thrust interference,

4) rheology and structural style,

5) kinematic indicators and

6) structural cross-section construction.

2.1 THRUST GEOMETRIES.

The theory of thrust tectonics has been slowly refined since its introduction earlier this century by geologists such as Argand (1916, 1924) and Rich (1934). A summary of the basic principles follows.

A thrust is a contractional fault (Norris 1958) that shortens datum surfaces such as bedding. Rocks above the thrust plane lie in its "hangingwall" and those below the thrust lie in its "footwall". During compression, anisotropies within the rock sequence e.g. sedimentary layering, metamorphic foliation are exploited as planes of weakness to form thrust "flats". Resistance to slip on this surface may cause the thrust plane to steepen and rise to structurally higher plane of weakness. This "steep" (Douglas 1950) is termed a thrust "ramp". A thrust surface therefore displays a "staircase trajectory" comprising flats and ramps (Butler 1982). Typically a flat will be bedding-parallel in sedimentary cover sequences and ramps will truncate bedding at low "cut-off angles" of 15-35° (Fig.2.3).

If a ramp is oriented perpendicular to the thrust movement direction then it is called a "frontal" ramp and will be characterised by dip-slip movement. A "lateral" ramp lies parallel to the movement direction and displays strike-slip movement. Oblique orientations also occur. Such footwall structures also are also reflected in the hangingwall as "culminations"

THRUST SYSTEM

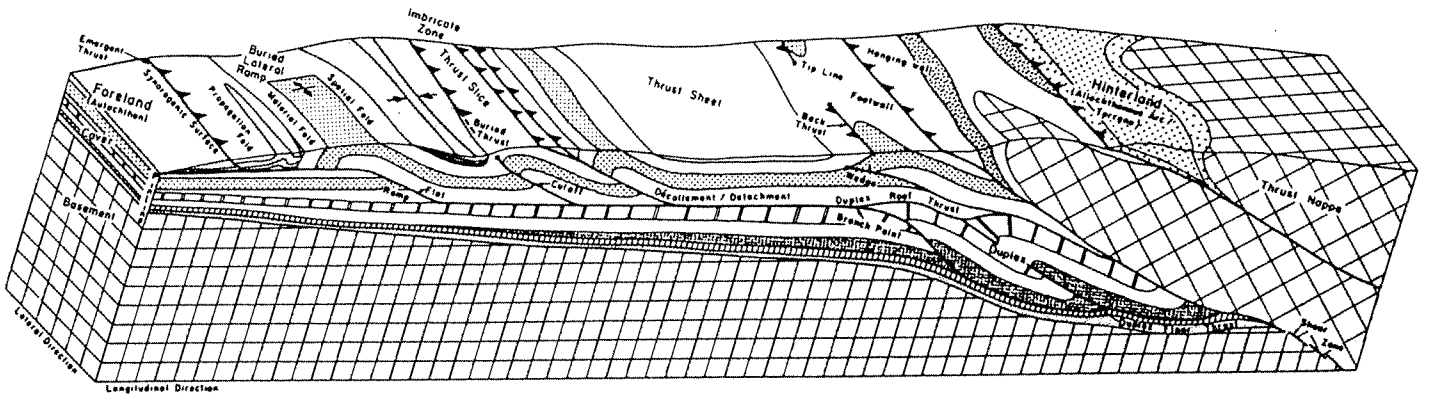


Fig.2.1. Block diagram illustrating the basic elements of a thrust system (from De Paor, 1988).

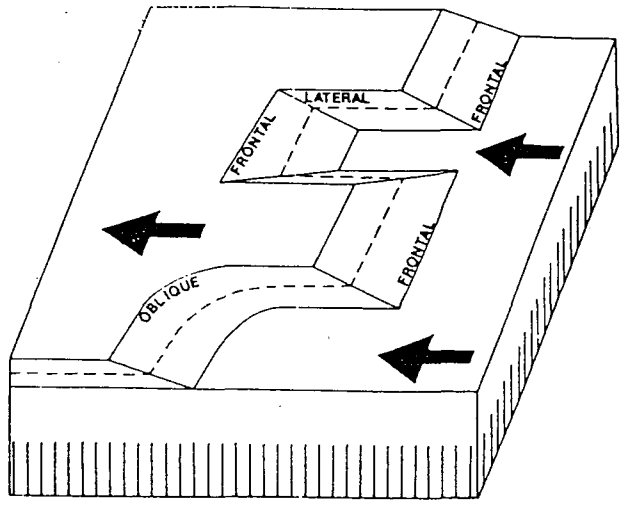


Fig. 3. Block diagram of footwall topography. The thrust transport direction is indicated by arrows.

Fig.2.2a. Diagram illustrating thrust footwall geometries.

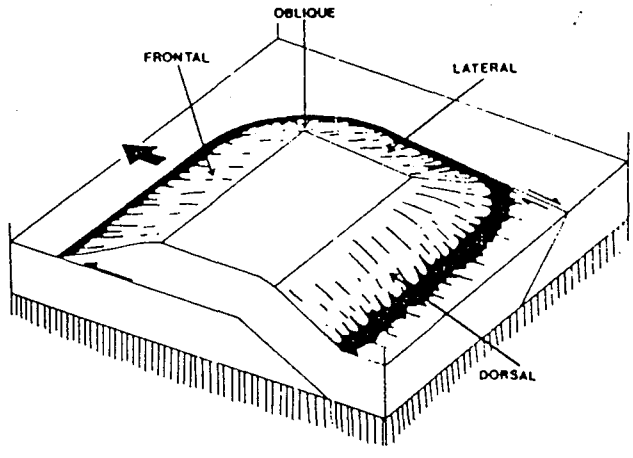


Fig. 11. Block diagram showing orientations of culmination walls relative to a thrust transport direction (large arrow).

Fig.2.2b. Diagram illustrating thrust hangingwall culminations.

(from Butler, 1982)

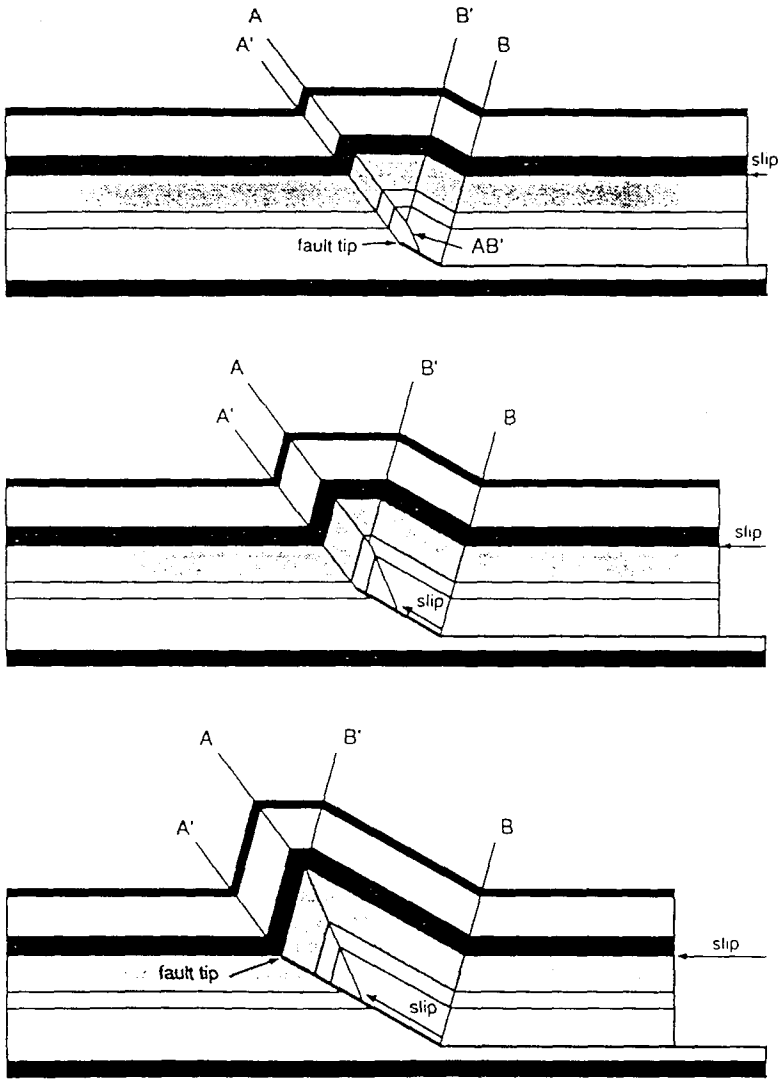


Fig.2.3. Fault-bend folding model.
(from Suppe & Medwedeff, 1990)

which are also classified as frontal, oblique and lateral (Fig.2.2a&b).

Generally a thrust will cut up-section in the transport direction. Exceptions occur however where a thrust breaches a pre-existing structure or stratigraphic variations.

Thrusts may propagate by footwall collapse to form a linked "piggy-back" system (Boyer & Elliott, 1982). Alternatively, initially isolated fractures may become linked as the deformation front advances. Breaching systems or "out-of-sequence" thrusts result from hanging-wall collapse (Fig.2.4). The timing and sequence of thrust movement is crucial to the understanding of a thrust system.

Thrusts are linked by "branch-lines", where thrust surfaces intersect, and these are represented on maps and sections by "branch-points". A thrust may have a single "trailing branch-point" e.g. in an imbricate stack, or where thrusts link again to a roof-thrust have a leading and a "trailing branch-point". Where imbricates have branch-points at the same stratigraphic horizon then a "duplex" (Boyer & Elliott, 1982) is produced (Fig.2.5). Duplexes may either have an "active roof-thrust" in which the strata in the hangingwall of the duplex are carried towards the foreland, or a "passive roof-thrust" in which strata in the hangingwall remain almost in situ as the duplex is wedged underneath it (Vann et al, 1986).

Displacment along a thrust decreases towards the "tip", where bed separation is zero and strain is accommodated by ductile folding (Fig.2.6). Such faults are termed "blind". Alternatively, a thrust may become "emergent" and transfer displacement across the topographic surface. At culminations and tipfolds out-of-syncline thrusts may nucleate in order to accommodate strain at the fold hinge (Fig.2.7). Such thrusts are linked to the thrust system by zones of ductile strain.

During thrust wedge development structures predominantly verge towards the foreland, but where friction at the base of the wedge is low a more symmetrical structural style

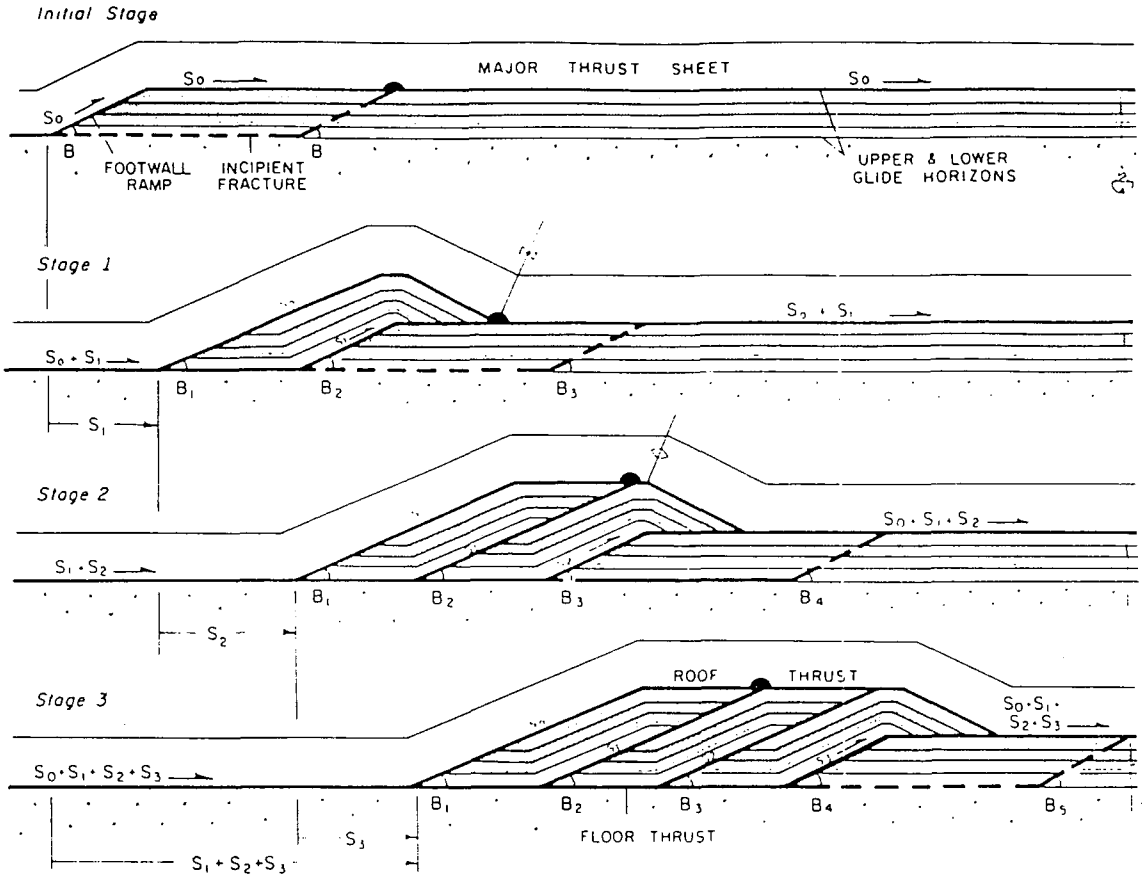


Fig.2.5. Diagram illustrating duplex formation. (from Boyer & Elliott, 1982)

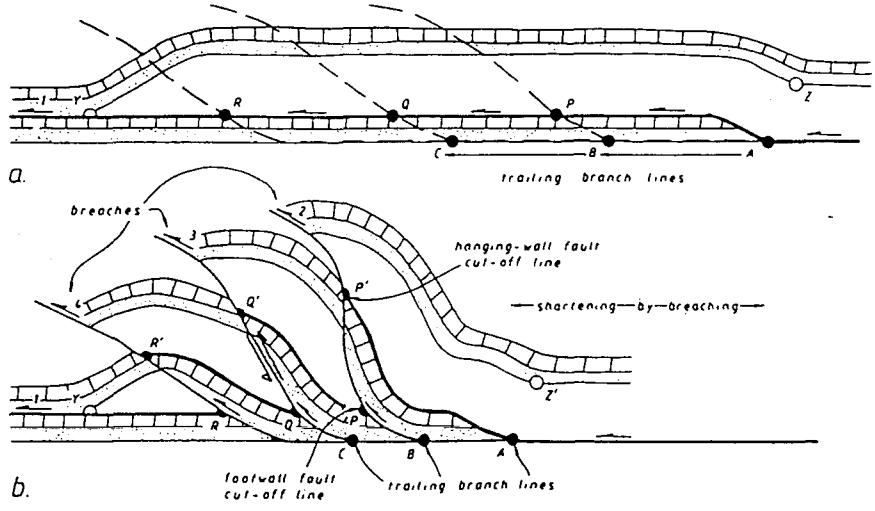


Fig. 10. The generation of breached roof geometries. (a) Restored section; (b) deformed balanced section (based on Butler 1984b). Trailing branch lines of the breaching imbricate thrusts are lettered A–C in their development sequence. Fault cut-offs (P, Q, R) are offset (R–R' etc.) with displacement so that the upper thrust sheet is shortened after its emplacement (e.g. length Y–Z' compared with Y–Z). The thrusts are numbered in their movement sequence (1, oldest; 4, youngest).

Fig.2.4. Diagram illustrating breaching thrust geometry. (from Butler, 1987)

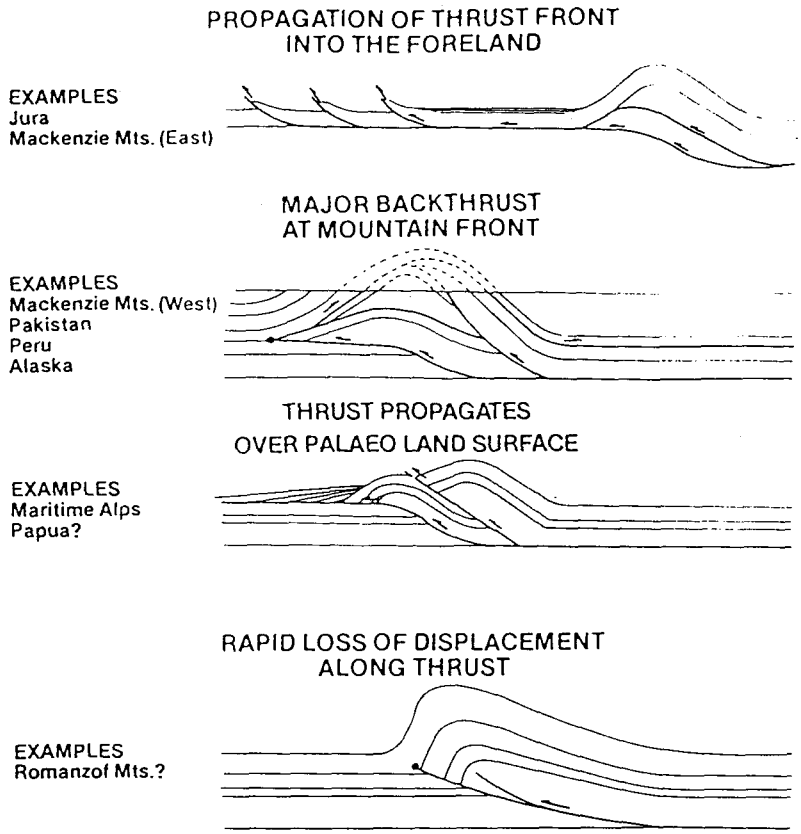


Fig.2.6. Styles of fault tip.
(from Vann et al, 1986)

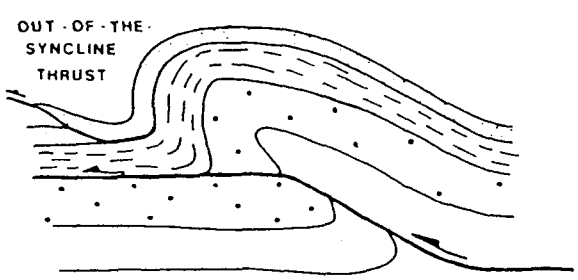


Fig. 14. An out-of-the-syncline thrust.

Fig.2.7. Out–of–syncline thrusting.
(from Butler, 1982)

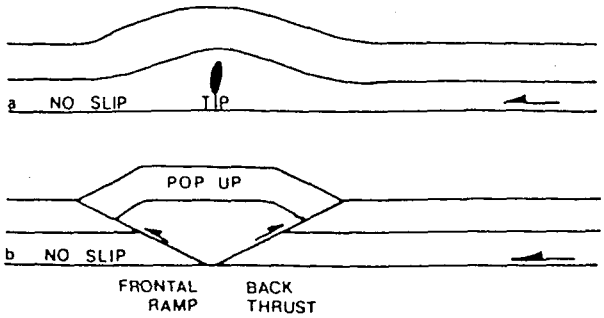


Fig. 16. Development of a pop up. (a) Layer-parallel shortening at a tip in front of a propagating thrust. (b) The developed pop up after thrusting.

Fig.2.8a. Pop–up formation

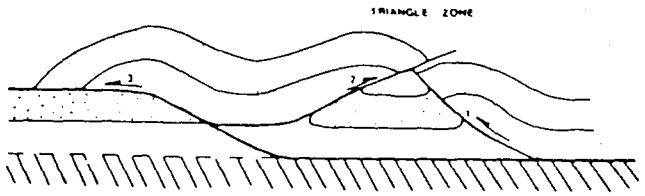


Fig. 17. Development of a triangle zone where a pop up back thrust (2) and a forward-directed imbricate thrust (1) converge. The thrusts are numbered in order of their relative displacement and the sequence continues with the development of a frontal ramp (3) at the leading edge of the pop up.

Fig.2.8b. Triangle zone formation

(from Butler, 1982)

results and back- and fore-thrusting play more equal roles. The result is the formation of "pop-up" and "triangle" zones (Fig.2.8a&b).

This style of thin-skinned thrusting is common in foreland regions where the undeformed foreland was characterised by "layer-cake" stratigraphy. Often the thrust belt foreland was an extended passive margin prior to collision. The inherited basement topography, structures and sedimentary cover sequences of such margins tend to deform in a more thick-skinned style. In such belts, the thrust geometry is controlled to a large extent by the extensional heritage resulting in inversion geometries (Gillcrist et al, 1987) and deep-seated control on cover thrust sequences (Roure et al, 1990).

2.2 THRUST WEDGE PROPAGATION THEORY.

Early studies of mountain belts imply that thrust belts are wedge-shaped, tapering to point towards the foreland (Argand, 1916). Recent studies of active or young thrust belts have permitted constraint of the thrust wedge geometry during orogeny because their topographic expression and basal detachment geometry have not been altered significantly by erosion and uplift due to isostatic compensation.

The inherent assumption that thrust belts could be regarded as wedges was modelled by Chapple (1978). He recognised four features, common to all thrust belts, that must be satisfied by any mechanical model.

1) Deformation within the thrust belt was essentially thin-skinned.

2) The basal detachment of the wedge has low shear strength.

3) The sedimentary sequences involved in thrust belts tend to be wedge-shaped prior to deformation (often being continental passive margin sequences) and this thickening is emphasised by the deformation.

4) The wedge material is shortened and thickened during deformation.

Thrust wedge analysis was advanced during the 1980's (Davis, Suppe & Dahlen,

1983; Dahlen, Suppe & Davis, 1984; Davis & Engelder, 1985) by the concept of the "critical wedge", which contends that a wedge must attain a critical taper in order to move on its basal detachment without requiring internal failure. The critical taper is dependant upon the slope of the basal detachment, the topographic slope (together defining the wedge taper), the internal strength of the wedge and the strength of the basal detachment. A subcritically–tapered wedge can not move upon its basal detachment without internal thickening and shortening. A critically– or supercritically–tapered wedge can glide on its basal detachment without internal deformation. However, the accretion of fresh material to the tip of the wedge causes a decrease in the wedge taper – a process that takes the wedge into the subcritical field. An accreting wedge must therefore undergo changes in order to remain supercritical. These include the following:

1) Internal deformation to thicken the wedge. Three major processes can contribute to the internal thickening of a thrust wedge. Backthrusting and backfolding, out–of–sequence thrusting and penetrative strain such as cleavage development and pressure solution (Geiser 1988) can substantially uplift, shorten and thicken the thrust wedge. However from published field examples these processes do not contribute sufficient thickening to satisfy theoretical models (Woodward 1987).

2) The addition of material to the wedge by underplating. The incorporation of new material into the wedge by this process both thickens the wedge and increases the basal slope and hence the wedge taper. It is therefore a particularly effective way of maintaining a supercritical taper. Although this process is clearly important, it is difficult to constrain in the field. It may have a major role in the uplift of high pressure lower crustal rocks in the internal zones of thrust belts (Platt, 1986).

3) An increase in the internal strength of the wedge. Since the critical taper is proportional to the internal strength of the wedge any increase in the internal wedge strength will permit the wedge to propagate at a lower taper. Wedge strength is related, not to the yield strength of the rocks comprising the wedge, but to the yield strength of anisotropies within the wedge i.e. bed–

ding planes, foliations, fault planes, etc.. The wedge is weakest when these anisotropies are oriented favourably with respect to the stress field i.e. at a low angle to the principal stress direction. Back-steepening of an imbricate stack or backfolding will therefore effectively strengthen the wedge, pushing the wedge towards a more supercritical state.

4) A decrease in the basal strength of the wedge. This will also permit wedge propagation at a lower critical taper. Grain size reduction during the formation of mylonites dramatically reduces shear strength (White & Knipe, 1978; White et al, 1980) and the exploitation of weak horizons such as evaporites (Davis & Engelder, 1985) are both processes that facilitate easy-slip.

It is clear that a combination of these processes takes place in order to maintain a supercritical state as an accreting thrust wedge propagates. However the downward flexure of the footwall to the thrust wedge due to the thrust load is the primary factor permitting continued wedge propagation. The advance of the thrust wedge causes the subsidence of the foreland resulting in the deposition of thick synorogenic sedimentary sequences. The flexure of the underriding footwall plate generates the basal slope of the wedge – a slope that increases towards the rear of the wedge, so increasing the wedge taper and promoting a supercritical state. It also permits the continued exploitation of easy-slip detachment horizons, such as the basement-cover interface. Without such subsidence, the basal detachment of the wedge would have to continually downcut into lower horizons in order to accommodate further tip accretion, since gravitational body forces would otherwise arrest overthrusting.

Later in the orogenic phase isostatic compensation causes the uplift of the internal zones of the mountain belt, so changing the orientation of structures and causing extensive unroofing and erosion of the thrust wedge. This makes the study of active thrust wedges vital to the true understanding of thrust wedge mechanics.

2.3 THRUST INTERFERENCE.

The basic linked framework of a thrust system was described earlier. Here the disruption of such a thrust system by the superposition of later compressive phases, whether related to the same orogeny or a much later event, is discussed.

The application of a compressive stress to a rock system will produce failure when the yield strength of part of that rock system, whether the rock itself or a pre-existing plane of weakness within the rock, is exceeded (Jaeger & Cook, 1979). Pre-existing fault planes may yield before the rock itself depending upon their orientation with respect to the stress system imposed. From the Mohr diagram it is clear that the failure criteria of the pre-existing fault may often be reached, for a given differential stress and shear stress, before that of the rock. Thus during the superposition of a compressive phase upon a previously deformed region one expects to find examples of both thrust reactivation and breaching thrusts.

Breaching systems occur in systems where backsteepening of the thrust pile has occurred, either as a result of piggy-back thrusting or backfolding. Where bedding is steeply-dipping both bedding planes and bedding-parallel thrusts are unfavourably oriented for reactivation and under colinear conditions breaching systems develop. Under non-colinear conditions either breaching systems or strike-slip reactivation may occur. However, if backfolding is present then even strike-slip reactivation is prevented and the region tends to become corrugated.

If the superimposed phases are colinear, then the reactivation of structures will be in the same sense as during the initial phase; frontal ramps will exhibit dip-slip in both phases, lateral ramps will exhibit strike-slip in both phases. However, when the superimposed phases are non-colinear then early lateral ramps may be reactivated as frontal ramps or backthrusts, and frontal structures may reactivate as laterals, etc.. Structures that may be too steep for reactivation under stresses normal to the fault plane may be easily reactivated as strike-slip or oblique-slip

faults under non-colinear stress regimes (Gillcrist 1988). Poorly-oriented early structures will be folded or breached by shallowly-dipping thrusts.

So far it has been assumed that the pre-existing structures have been compressional. If the pre-orogenic foreland was a continental passive margin, extensional faults will have generated a series of horsts, graben and half-graben. The inherited stratigraphy can not then be accurately modelled as a "layer-cake", rather the model must account for a basement-cover contact that is offset by faults and a sedimentary cover consisting of pre-, syn- and post-rift sequences. The reactivation of such a region is complex. If the extensional faults are shallowly- to moderately-dipping (Price & Todd, 1988) they may be reactivated. Where the faults are steeper they resist reactivation and act as buttresses to compressional breaching thrusts, producing a variety of "short-cut" geometries (Gillcrist et al, 1987).

2.4 RHEOLOGY AND STRUCTURAL STYLE.

The structural style of deformation is scale-independent but depends upon the physical properties of the rock units involved and the physical conditions under which they deformed. For a given material the structural style may be characterised by folding whereas, under different stress or strain rate conditions brittle failure may have resulted. Likewise, under certain physical conditions a shale may approximate to plastic behaviour whilst a sandstone may remain unstrained.

For a single lithology two physical properties are of primary importance - the compressive strength and the shear strength. In addition, the presence of planes of weakness within the rock e.g. bedding planes, fracture planes, etc. will reduce the strength of the rock. Since rocks are normally anisotropic materials the orientation of the rock with respect to the stress field is also important. In sedimentary sequences, limestones and coarse-grained clastics e.g. sandstones and conglomerates are normally "strong" or "competent", whilst fine-grained clastics and evaporites are normally "weak" or "incompetent". Competent units are able to with-

stand greater differential stress conditions than incompetent ones, although competent units may be rendered weak by fracturing or high fluid pressures.

Whether failure occurs by faulting or folding, in a given rock, depends largely upon the constraining stress (and hence the differential stress), the strain rate and the orientation of rock anisotropies with respect to the stress field. The initiation of buckling within a single layer depends upon the thickness of the layer and the competence contrast between the layer and the surrounding matrix. A thick layer with low competence contrast requires a greater stress to deform it than a thin layer with high competence with respect to its matrix. However, in nature one is confronted by a rock sequence consisting of a number of such "single layers" whose strain fields interfere with each other during deformation and hence it is necessary to consider the sequence as a whole (or multilayer), rather than its individual members (Price & Cosgrove, 1990).

Molasse sediments, comprising alternating incompetent shales and competent sands, typify such a multilayer. In the pure shear regime, multilayer buckling theory predicts kinkband folding or tangential, longitudinal strain buckling depending upon the relative values of the normal and shear stress coefficients and the applied stress (*op.cit.*). In a simple shear regime the much weaker incompetent beds are exploited, resulting in a distributed simple shear deformation within the incompetent units with only slight straining of the competent units – similar to a deck of cards.

In a multilayer with low competence contrast, the behaviour of the sequence tends to that of a single, thick layer and hence large scale buckling or faulting occurs.

In the external zones of fold–and–thrust belts, layer–parallel shortening (LPS) and buckling occur in the early phases of deformation (Ford, 1987). As the basal decollement propagates into the foreland the material in its hangingwall is compressed, but the low angle between the bedding anisotropies and the principal stress direction favours pure shear deforma–

tion processes such as LPS and buckling.

From a structural viewpoint, the stratigraphy of a region can often usefully be regarded in terms of its rheological properties. Such mechanostratigraphic analysis is crucial to the understanding of variations in structural style and can then be used to predict stratigraphic variations where the structure style is understood.

2.5 KINEMATIC INDICATORS.

Describing the role of transport direction variations in controlling structural style within mountain belts is one of the main objectives of this thesis. The accurate determination of palaeo–transport directions is therefore crucial to the project. Data on all scales, from microscopic to regional, provide such constraints on transport directions and in this section the criteria that have been used are presented.

The majority of the transport directions obtained were derived from the analysis of shear zones at outcrop scale. The transport direction is indicated by three distinct types of lineation within a shear zone:

1) stretching lineation – produced by the ductile shearing of minerals during deformation so as to yield elongate ribbons with a preferred alignment parallel to the long axis of the strain ellipse (or transport direction).

2) fibrous lineation – produced by the precipitation of minerals, commonly calcite or quartz, on the fault plane during deformation. Since the fibres grow under stressed conditions they are aligned in the shear direction and the sense of shear can also be deduced.

3) striations – where no mineralisation has occurred but the abrasion and scouring of the rock surfaces during shearing produced lineations parallel to the transport direction.

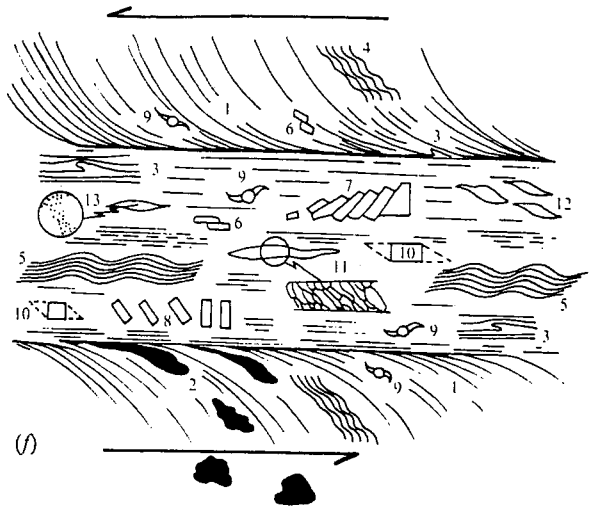
Although lineations are accurate transport direction indicators, the sense of shear

must often be deduced from other features of the shear zone. The best shear sense indicator is the deflection of pre–existing foliations or bedding into parallelism with the shear zone boundaries (Ramsay, 1980) or the offset of marker horizons. If the margins of the shear zone are not exposed then a number of features within the shear zone, if taken in conjunction, will define the shear sense (Fig.2.9). The relationship of composite planar fabrics (s– and c–foliations) can be confidently used, as can the presence of shear bands (the extensional crenulation cleavage of Platt & Vissers 1980). Intrafolial folds, augen structures (Simpson & Schmid 1983) or tailed rotating clasts (White, Bretan & Rutter 1986) provide a clear indication of vorticity but the interpretation of extensionally fractured clasts and preferred crystallographic orientations requires greater care.

Under brittle conditions, brittle "R" (Riedel) and "Y" shears will develop (Ramsay & Huber, 1987; White, Bretan & Rutter, 1986), which again yield reliable shear sense data.

On a regional scale, the map traces of faults and fold axes can be used to constrain the transport direction. The interpretation of the three dimensional structure of thrust belts from map patterns, using fault tiplines, branchlines and cut–offs has been described by Diegel (1986). The identification of lateral structures is particularly useful in constraining the tectonic transport direction. Fault–propagation folds trend approximately perpendicular to the transport direction and their sense of vergence usually represents the transport direction. However, as in the study of shear zones, it is necessary to assess further data to ensure that an accurate shear sense is deduced – in many external fold–and–thrust belts backfolding and backthrusting is well–developed, particularly in thrust wedges underlain by evaporites.

The study of cleavages is a further constraint upon transport direction. Cleavage develops in the XY–plane of the strain ellipsoid (Ramsay, 1980) and is often related to fold generation. The bedding–cleavage intersection lineation can therefore be used in the same way as fold axial data.



(1) rotation of a pre-existing or generated foliation:
(2) rotation of deformed markers:
(3) asymmetry of intrafolial folds;
(4) normal kink-bands (microshears) in the margin or
(5) central fabric of the shear zone:
(6) sheared porphyroclasts;
(7) rotation of fragments owing to shear fractures;
(8) rotation of fragments owing to tensile fractures;
(9) asymmetry of trails growing around rotating clasts;
(10) asymmetry of trails growing around non-rotating clasts;
(11) asymmetry of elongate recrystallised quartz grains;
(12) asymmetry of dragged-out mica porphyroclasts;
(13) asymmetry of quartz c-axia fabrics.

((*f*) after White *et al.*, 1986).

Hence, a combination of small-scale and large-scale observations can be used to constrain the tectonic transport direction. Once this direction is ascertained and any subsequent rotations of that kinematic indicator are known, meaningful cross-sections can be constructed and interpretations of the regional geology made.

2.6 STRUCTURAL CROSS-SECTION CONSTRUCTION.

The use of the balanced cross-section in the analysis of fold-and-thrust belts has become widespread since the mid-1960's when the technique, developed in the Appalachian and Rocky Mountains, was applied to the classical Moinian and Alpine belts. Although the principles of cross-section construction are implicit in the work of Argand (1924), Rich (1934) and others, they were first explicitly described by Dahlstrom (1969).

Dahlstrom argued that in any given region where the structural style of deformation was known, a cross-section that geometrically balanced would be a more tightly constrained solution than one that did not. His arguament was based upon the principle of the conservation of volume - a principle that could be applied to the Alberta Foothills. By assuming a concentric style of deformation and constant bed thickness, the three dimensional "volume conservation" could be reduced to a one dimensional "line-length conservation". Hence, Dahlstrom proposed the line-length balancing technique in which all deformed horizons could be restored to a layer-cake, undeformed template with beds having equal lengths in both deformed and restored sections. If a cross-section could be balanced it was deemed permissible although not necessarily correct.

A section is further constrained by the need for consistency with sections constructed parallel to it. In the regime of concentric deformation the amount of shortening exhibited by adjacent sections should be similar because orogenic shortening along a mountain belt varies very gradually, although the shortening accommodated on individual structures may vary quite rapidly. With sufficient data palinspastic maps can be drawn, further constraining the permissible cross-sections.

It is emphasised that section balancing is simply a geometrical constraint additional to the geological constraints appied by all the available data. Elliott (1983) suggests that a section is "viable" if it can be restored and "admissible" if it incorporates all available data and reflects the observed structural style. A "balanced" section is one that is both viable and admissible.

Line-length balancing is particularly applicable to the external zones of mountain belts where the majority of shortening is accommodated on thrust planes and the strain values within thrust sheets are low. Precise theories have been generated to model this situation, e.g. Suppe (1983). This model assumes that all observed folding is in response to movement over thrust ramps and that such fault-bend folds can be approximated by kink domains. Hence, straight bedding panels are separated by sharp, angular fold hinges.

An alternative method is to use the area balancing technique. Here the restored length of a thrust sheet can be estimated from its deformed area and its predeformational thickness allowing the rapid restoration of zones of particular complexity or of minor importance. The method can also be used interpretatively to deduce the depth of detachment and percentage shortening in the deformed wedge (Hossack, 1979).

The introduction of volume loss invalidates both line-length and areal balancing approaches to cross-section restoration. Hossack (1979) supports the use of balancing methods even where volume loss is known, on the grounds that section balancing will always yield a minimum estimate of tectonic shortening. Volume loss can occur due to sedimentary compaction during burial, tectonic compaction during deformation, pressure solution processes or flattening strain, but the extent of such processes is difficult to constrain. Hossack (1979) cites approximately 10% volume loss (equivalent to approximately 6.7% cross-sectional area loss) for one example. Woodward et al (1986) suggest individual restoration of external zone imbricates to incorporate strain and cleavage, line-length balancing being used for competent, uncleaved

horizons and area balancing being used for incompetent, strained units.

The recognition of strain in the traditionally "unstrained" external zones of mountain belts has led to the construction of more representative cross-sections. Cooper et al (1986) and Geiser (1988) describe the role of layer parallel shortening both in the early and later stages of thrust wedge propagation. Ford (1987) presents a sequential balancing technique that accounts for the changing style of deformation throughout mountain building.

Hence, the balancing technique applied in cross-section construction must be selected to account for differing deformation mechanisms both through time and in different rock rheologies.

Balanced cross-sections constructed across the Appalachian and Rocky Mountains (Bally, Gordy & Stewart, 1966) restore back to a layercake stratigraphic template, depicting a simple predeformational depositional setting in which sedimentary thickness variations were minor. Many mountain belts result from the collision of previously extended continental passive margins and hence, the restored stratigraphic template for such regions should reflect the rifted, extensional regime that existed prior to orogeny. The western Alps represent a classic example of such a region, the French foreland having been rifted prior to the opening of the Ligurian Tethys in early Mesozoic times before undergoing strong inversion and orogeny during the Tertiary period. The strong stratigraphic variation exhibited in the Alps had long been used as the basis of Alpine zonation (Trumpy, 1960; Debelmas & Lemoine, 1970), but the tectonic evolution of the region has only been described in recent times (Tricart, 1984; Tricart & Lemoine, 1986; Lemoine et al, 1986). The appreciation of the role of inherited structure in dictating the structural style of orogenic shortening leads to the construction of very different "balanced" structural cross-sections (Butler, 1983; Butler, Matthews & Parish, 1986; Gillcrist, 1988; Gillcrist, Coward & Mugnier, 1987).

Balancing of cross-sections should be undertaken to geometrically constrain

sections, but this should not be done at the expense of known geological data.

CHAPTER 3.

AN INTRODUCTION TO THE WESTERN ALPS.

The Alpine chain has been the object of geological study for over 150 years and in that time a vast amount of literature has accumulated, making the Alps the most analysed orogenic belt in the world. This chapter outlines the following aspects of Alpine geology :-

1) Tethys; from geosynclinal to plate tectonic perspectives.

2) Passive margin development in SE. France; the recognition of pre-, syn- and post-rift sequences.

3) The stratigraphy of the Dauphinois zone, SE. France; continental rift, marginal sag and intermontane basin deposits.

4) Structural interpretations of the western Alps; the role of Variscan and Tethyan inherited structures, thick- and thin-skinned Alpine deformation, Alpine inversion of Tethyan basins and phases of Alpine deformation (U.Cretaceous to Present).

3.1. ALPINE RESEARCH AND THE APPLICATION OF PLATE TECTONICS.

Among the most important early works were those of Suess (1904, 1908, 1909, 1924) in which the existence of Tethys, a Mesozoic oceanic realm lying between Europe and Africa, was proposed. The presence of ophiolitic remnants within the Alpine chain testifies to the existence of such an ocean during the Jurassic and Cretaceous periods, the closure of which coincided with the development of the Alps. Two opposing schools of thought emerged to explain the closure of the Tethyan Ocean during the early 20th Century - the "**fixists**" who invoked vertical movements, and the "**mobilists**" (Wegener, 1912, 1929; Argand, 1924) who proposed that the continents were not constrained to vertical movements but had "drifted" across the earth's surface through time.

Debate continued until a mechanism to account for the horizontal movement of continents

was proposed. This came in the 1960's in the form of **the theory of plate tectonics** which combined the geological and geophysical evidence of continental drift and seafloor spreading (Vine, 1966; Le Pichon, 1968; Dewey & Bird, 1970).

Recent plate kinematic studies view the evolution of the Alpine chain from the perspective of the relative motion of Africa with respect to Europe since Triassic times (Dewey et al, 1973; Biju-Duval et al, 1977; Zonenshain et al, 1986; Dercourt et al, 1986; Le Pichon et al, 1988; Dewey et al, 1989). The relative continental motion of Africa and Europe is calculated indirectly by using the spreading histories of the Central and North Atlantic Oceans (Bullard et al, 1965; Smith, 1971; Pitman & Talwani, 1972; Dewey et al, 1989; Helman, 1990). Relative plate motion vectors do not account for the range of thrust transport directions within the Alpine chain due to the role of independent microplates, e.g. Apulia, and crustal heterogeneities. However, Dewey et al (1989) identified six phases of Europe-Africa motion since the opening of the Central Atlantic which are in broad agreement with the observed Alpine geological data (Fig.3.1).

1) Opening of the Central Atlantic (Callovian-Aptian). Continental rifting between Africa and N.America characterised the Triassic-Liassic period. Break-up did not occurr until the Callovian (Hiscott et al, in press) or slightly earlier (175Ma; Dewey et al, 1989) (Fig.3.1) and from then until Anomaly M-0 (118Ma) spreading in the Central Atlantic was bounded in the north by the Azores-Gibraltar Fracture Zone. This sinistral transfer linked Central Atlantic spreading to the Ligurian Tethys. This period is correlated with the post-rift sequence of south-eastern France (Lemoine et al, 1986).

2) Opening of the North Atlantic from Iberia to Labrador (Aptian-Campanian). The initiation of spreading north of the Azores-Gibraltar F.Z. at *ca.* Anomaly M-0 (Group Galice, 1979) brought this extensional regime to an end. The locus of spreading in the North Atlantic migrated northward between Anomalies M-0 and 33 (118-74Ma) causing :-

(i) the rotation of Iberia about a pole close to that of Africa (Savostin et al, 1986; Dercourt et al, 1986),

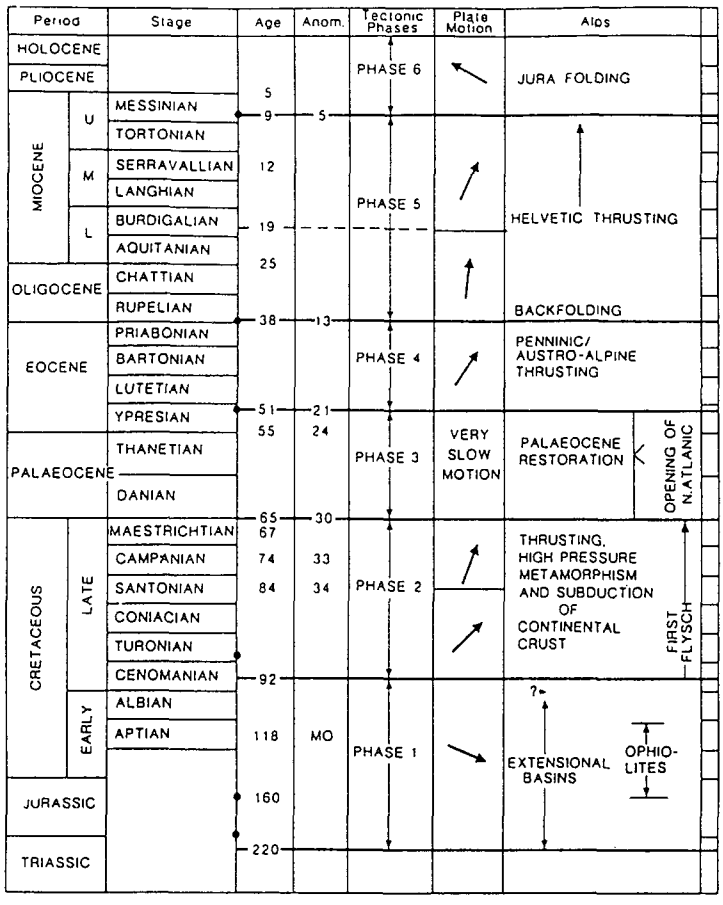


FIG. 2. Correlation chart showing the general direction of Africa's motion with respect to Europe and correlations with various tectonic events in the Alps, Mediterranean Sea, and the Atlantic Ocean.

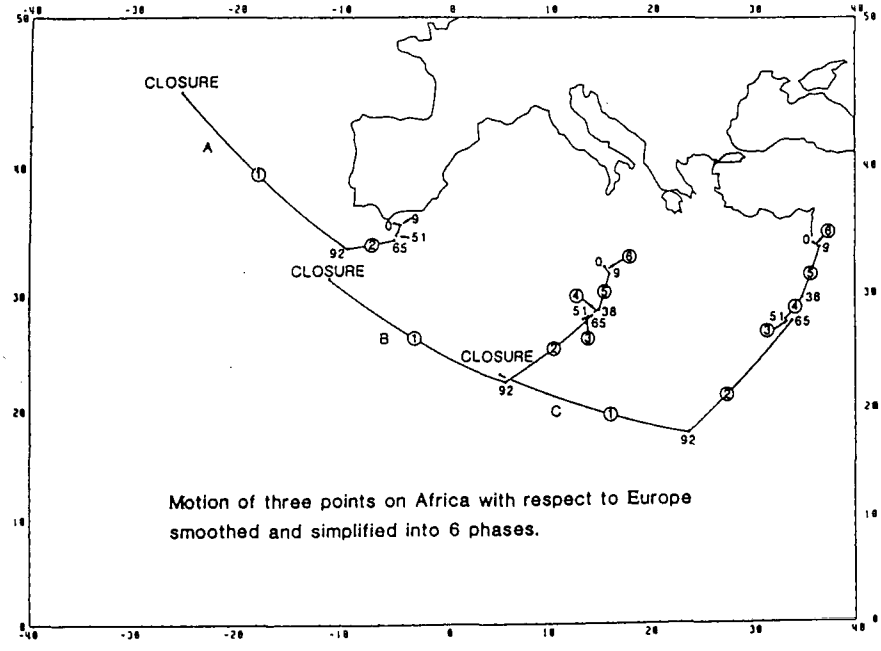


FIG. 3. Flow lines depicting the path of Africa's motion for the same three points as in Fig. 1 (a) after the motion path has been smoothed and corrected to correlate with the events and phases of motion shown in Fig. 2.

Fig.3.1. The Relationship between Europe/Africa Plate Motions and Alpine Events (from Dewey et al, 1989).

(ii) opening of the Bay of Biscay (Soler et al, 1981) and

(iii) abortive rifting in the Rockall Trough (Roberts, Masson & Miles, 1981).

During this period Africa moved northeastward with respect to Europe resulting in subduction, high pressure metamorphism in the eastern Alps and Pre-Senonian phase of folding (Flandrin, 1966; Siddans, 1979) and Helminthoid flysch deposition (Kerckhove, 1979) in the western Alps (Dewey et al, 1989).

3) Opening of the Labrador Sea (end Campanian-early Eocene). Between Anomalies 33 and 21 (74-49Ma), the relative motion between Europe and Africa was slow and erratic. However, during this period the Pyrénéan-Provençal phase of folding occurred in southern France.

4) Opening of the Greenland-Norwegian Sea (early Eocene-Present). Following Anomaly 21 (49Ma), spreading in the Labrador Sea ceased and the North Atlantic spreading ridge jumped to the east of Greenland. This resulted in the northward convergence of Africa with Europe and is associated with the main phase of Alpine overthrusting. This phase has been subdivided into Pre- and Post-Nummulitic events in the western Alps which are characterised by WSW- to WNW-vergent tectonic transport directions (Gillcrist et al, 1987; Platt et al, 1989). This has prompted Platt et al (1989) to propose that the NW-vergent motion of the Apulian microplate dictated thrust kinematics in the western Alps (Fig.3.2).

3.2. THE EVOLUTION OF THE SOUTHERN EUROPEAN MARGIN OF TETHYS.

Detailed stratigraphic investigations of the Western Alps have led to a good understanding of the evolution of this region since the Triassic. Classic reviews by Trumpy (1960) and Ramsay (1963) lay the foundation of Alpine zonation which is based on the recognition of five palaeogeographic domains (Fig.3.3). These are :-

1) the Helvetic-Dauphiné zone,

2) the Valais zone,

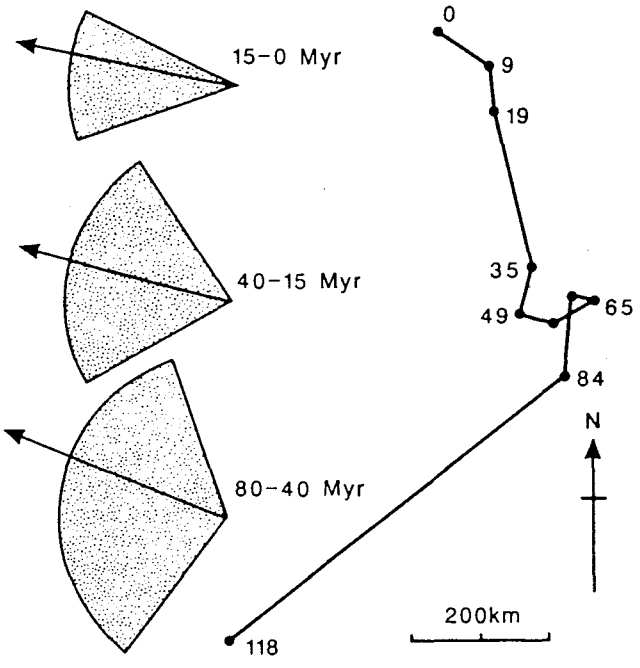


Fig. 2 Range of PMV orientations consistent with the kinematic data (shown as arcs), and our estimate of the most likely orientation (arrows), for times since the Late Cretaceous, compared with the Africa–Europe plate-motion path calculated for the location of Torino (Italy) from Africa–Europe poles redetermined by ref. 5.

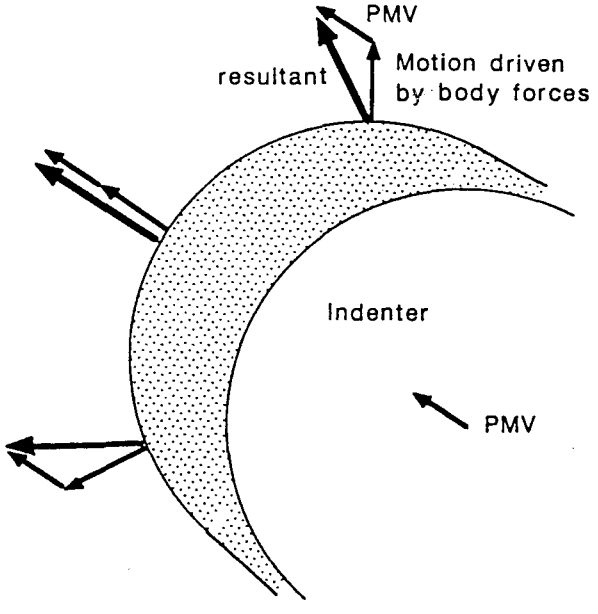


Fig. 3 Motion vectors produced by the interaction between the plate-motion vector (PMV) and body forces acting normal to the trend of a mountain belt; the latter should lie within 90° of the PMV

Fig.3.2. The Disparity between Alpine Kinematic Data and Plate Motion Vectors (from Platt et al, 1989).

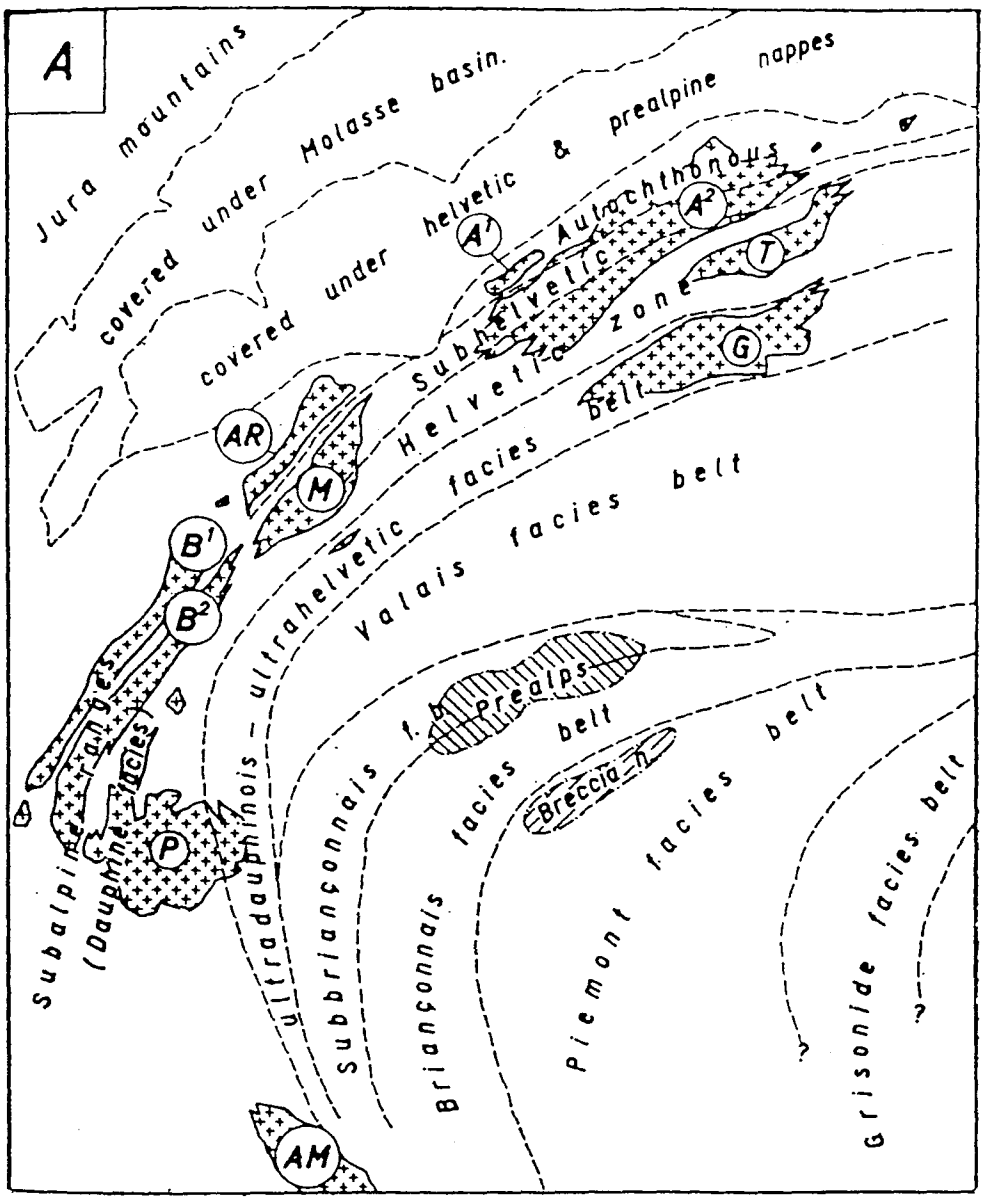


Fig.3.3. Classical Zonation of the Western Alps.

3) the Briançonnais zone,

4) the Piemont zone and

5) the Austro-Alpine zone.

During Tethyan evolution the Piemont zone became oceanic and separated the Dauphiné, Valais and Briançonnais zones of the European margin from the Austro-Alpine zone of the Apulian margin (Fig.3.4). In essence, these zones define platform regions (the Dauphiné, Briançonnais and Austro-Alpine zones) separated by deepwater basins (the Valais and Piemont zones).

The zones are characterised by different subsidence histories (Fig.3.5) which were originally explained in terms of geosynclinal theory (Trumpy, 1960) but have more recently been re-interpreted in terms of plate tectonics and passive margin development (Lemoine, 1975, 1983; Graciansky et al, 1979; Tricart & Lemoine, 1986). This concept is reviewed by Lemoine et al (1986) who discuss the stratigraphic and structural evolution of the continental margin of Tethys in terms of the pre-, syn- and post-rift phases of passive margin development. These authors also note the influence of Tethyan heritage upon the style of later Alpine compressional structures.

The stages of passive margin development in SE. France can be summarised as follows :-

1) The Hercynian metamorphic basement suite is unconformably overlain by U. Carboniferous to Permian continental deposits laid down in trough-like basins (Ramsay, 1963). These have been interpreted as molasse deposits (Tricart, 1984).

2) The Triassic heralded marine transgression across the eroded Hercynian landscape and the onset of Tethyan rifting. A thin, evaporitic, shallow-water sequence, including occassional volcanics, accumulated on the slowly subsiding Dauphiné zone (Debelmas & Kerckhove, 1980; Vernet, 1965) whilst in the rapidly subsiding Briançonnais zone a much thicker shelf carbonate sequence occurs (Megard-Galli & Baud, 1977). Triassic rifting, therefore, initiated the distinction between the Dauphiné and Briançonnais zones (Fig.3.4).

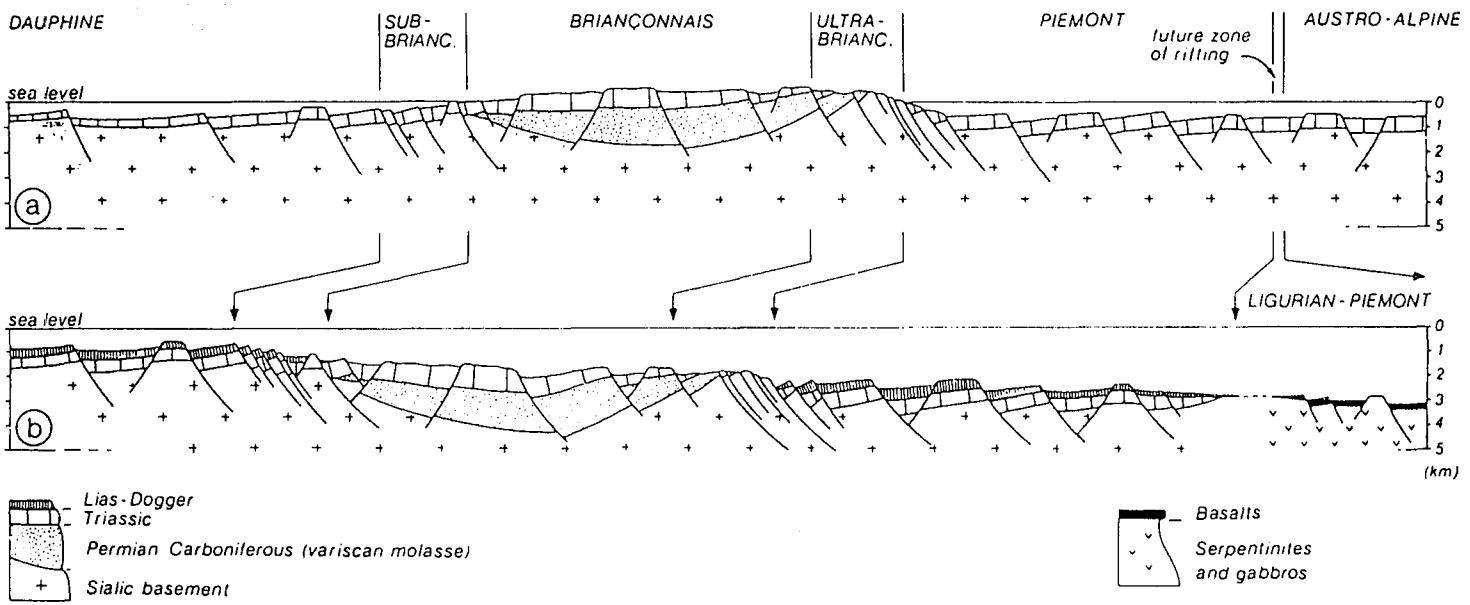


Fig. 3. Very schematic sections across the paleogeographic domains of the western Alps.

a. End of Lias, during the rifting of the Tethyan margins.

b. End of Malm, while the oceanic crust in the "Ligurian-Piémont domain" was spreading. Note the sinking of the whole European continental margin.

There is no exact horizontal scale; the total width represented could be about 500 km at least. Sketches modified after Bourbon and others (1975 and 1977), Graciansky and others (1979), and Graciansky and Lemoine (1980).

Fig.3.4. Cartoon of the Palaeogeographic Domains of the Western Alps (from Tricart, 1984).

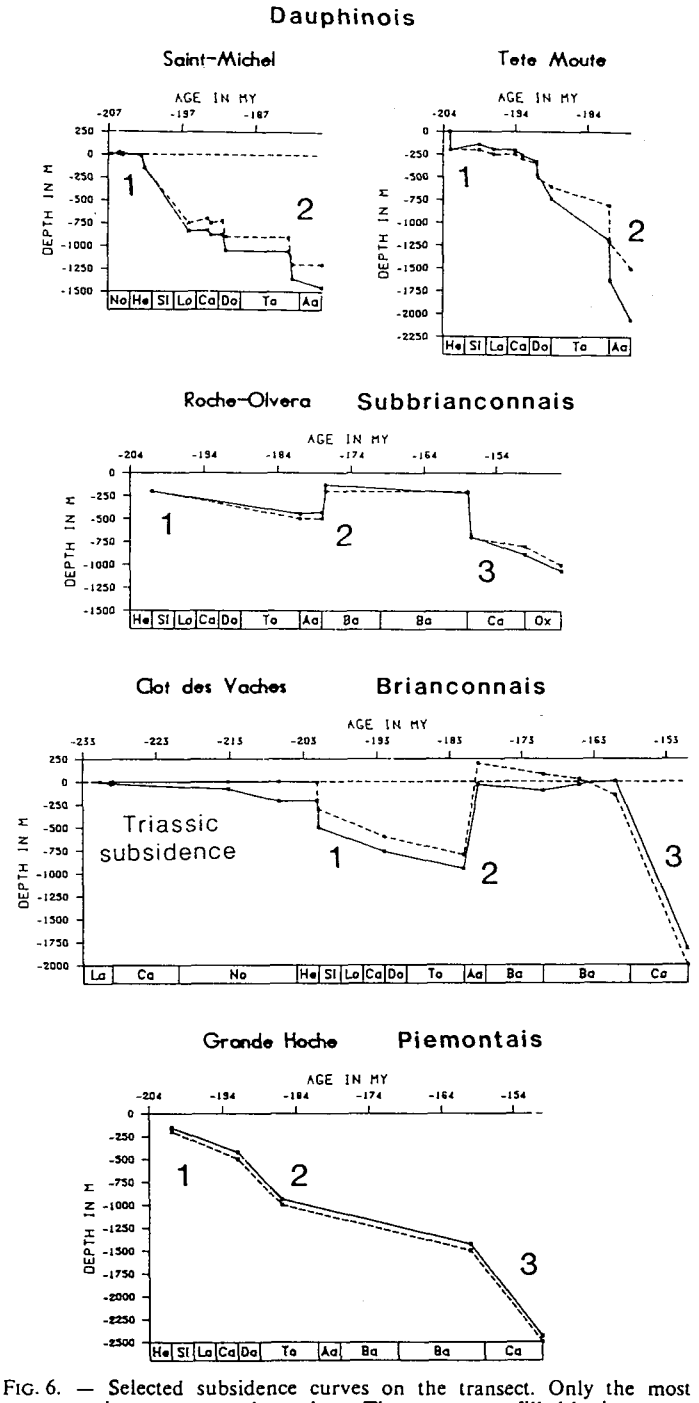


FIG. 6. — Selected subsidence curves on the transect. Only the most representative curves are shown here. These are water-filled basin curves, assuming local isostatic balance. The full line represents the tectonic subsidence, the dashed line the bathymetry. The numbers correspond to the main events. The vertical error on subsidence would lead, if represented, to a zone of 250 m width over and under each curve. Locations of sections, see figure 1.

Fig.3.5. Jurassic Subsidence Curves for the Palaeogeographic Domains of the European Margin, SE.France (from Rudkiewicz, 1988).

3) Major rifting occurred during the Liassic (Lemoine et al 1986). Major seafloor topography also developed at this time (Roux et al, 1988; Rudkiewicz, 1988), the Briançonnais zone experiencing uplift and emergence whilst the Dauphiné zone is characterised by basin-and-shoal deposition associated with half-graben formation (Barfety et al, 1979; Lemoine, Gidon & Barfety, 1981; Debelmas & Kerckhove, 1980) (Fig.3.6). Syn-rift Liassic sedimentation reaches 4km thickness in northern Pelvoux (Barfety et al, 1979; Grand, 1988) indicating major extension at this time (Fig.3.7).

4) By the Malm and Lower Cretaceous local fault-controlled depocentres had been replaced by regional subsidence. Subsidence rates exceeded sedimentation rates during this period and deepwater conditions were established in the Vocontian Trough/Dauphiné Basin (Lemoine et al, 1986; Rudkiewicz, 1988), typified by the "terres noires" facies. This foundering of the Dauphiné Basin and the Briançonnais platform is associated with the onset of oceanic spreading in the Ligurian Tethys. The Pelvoux massif, like the Provençal platform (Dardeau, 1988), may not have subsided at this time.

5) Renewed extension in the Lower Cretaceous is linked to rifting of the Bay of Biscay prior to spreading in the Aptian (Dardeau, 1988). During this period deepwater conditions persisted in much of the Dauphiné basin but during the Upper Cretaceous platform carbonates became widespread (Kerckhove, 1979). Clastic flysch of this age is found in the footwall to the Frontal Pennine Thrust indicating the incipient Alpine orogeny.

3.3. THE STRATIGRAPHY OF THE DAUPHINOIS ZONE.

The detailed stratigraphy of the western Alps has been documented by the Bureau de Recherches Géologiques et Minières (B.R.G.M.) and is summarised in their published maps at 1:50 000 and 1:250 000 scale. Here I provide only a brief résumé of the main features of the

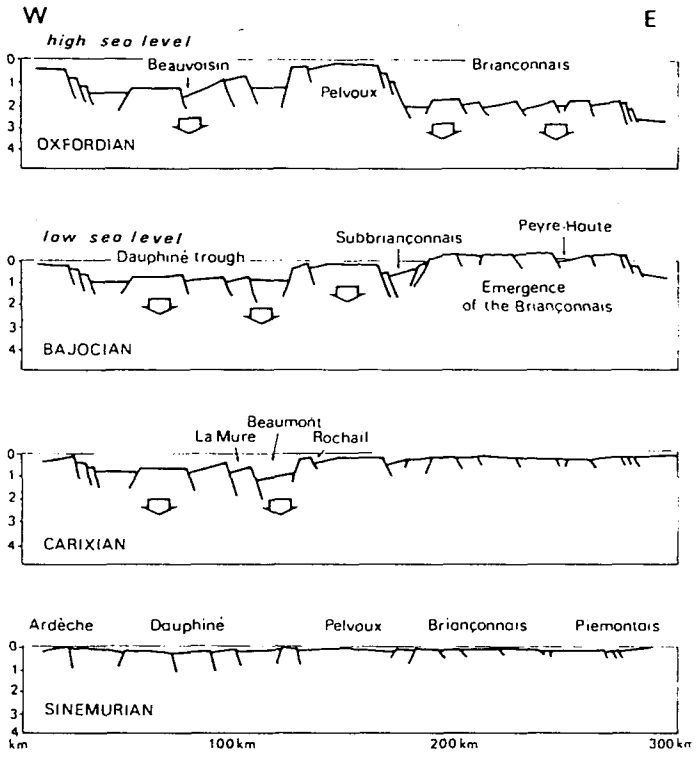


FIG. 8. — Reconstruction of bathymetric evolution of the Tethyan margir across the French Alps during Jurassic times.

Fig.3.6. Bathymetric Evolution of the Tethyan Margin in the Western Alps (from Roux et al, 1988).

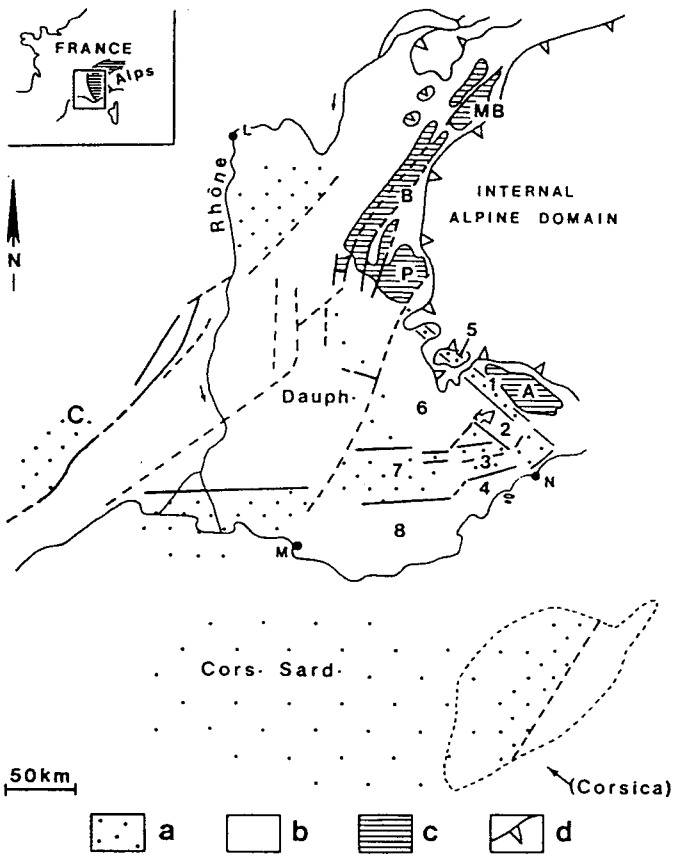


FIG. 2. — The Liassic and Middle Jurassic framework in the external alpine domain, according to Arnaud and Monleau [1979], Baudrimont and Dubois [1977], Jautée [1980], Lemoine [1985], Pairis [1965], and Plan [1964] (modified). — a : horsts or half-horsts : 1 : Tinée; 3 : Cheiron; 5 : Barcelonnette; 7 : Middle Verdon; C : « Cévenole » edge; (Cors. Sard.) Corsico-Sardinian block with a Jurassic « Cévenole » trend in Corsica [Lemoine, 1985], the distance from Corsica to Provence was possibly larger. — b : grabens or half-grabens : 2 : Barrot; 4 : Antibes; 6 : Gap-Digne; 8 : Provence; (Dauph.) Dauphiné. — c : external crystalline massifs : MB : Mont-Blanc; B : Belledonne; P : Pelvoux; A : Argentera. — d : front thrust of the internal alpine domain. — L : Lyon; M : Marseille; N : Nice.

FIG. 2. — *Organisation des structures du Lias-Dogger dans le domaine externe, d'après Arnaud et Monleau [1979], Baudrimont et Dubois [1977], Jautée [1980], Lemoine [1985], Pairis [1965] et Plan [1964] (modifié).* — a : *horsts ou demi-horsts :* 1 : *Tinée;* 3 : *Cheiron;* 5 : *Barcelonnette;* 7 : *Moyen-Verdon;* C : *bordure cévenole; (Cors. Sard.) bloc corso-sarde, avec un accident de direction cévenole en Corse durant le Jurassique [Lemoine, 1985], la distance entre Corse et Provence pouvant être plus importante.* — b : *grabens ou demi-grabens :* 2 : *Barrot;* 4 : *Antibes;* 6 : *Gap-Digne;* 8 : *provençal;* Dauph. *dauphinois.* — c : *massifs cristallins externes :* MB : *Mont-Blanc;* B : *Belledonne;* P : *Pelvoux;* A : *Argentera.* — d : *chevauchement frontal du domaine alpin interne.* — L : *Lyon;* M : *Marseille;* N : *Nice.*

Fig.3.6a. Basement Framework during Tethyan Rifting in the Western Alps (from Dardeau, 1988).

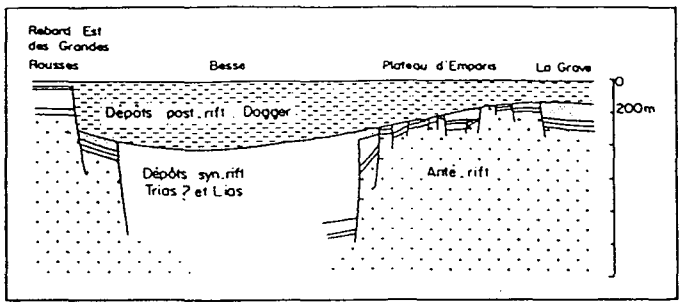
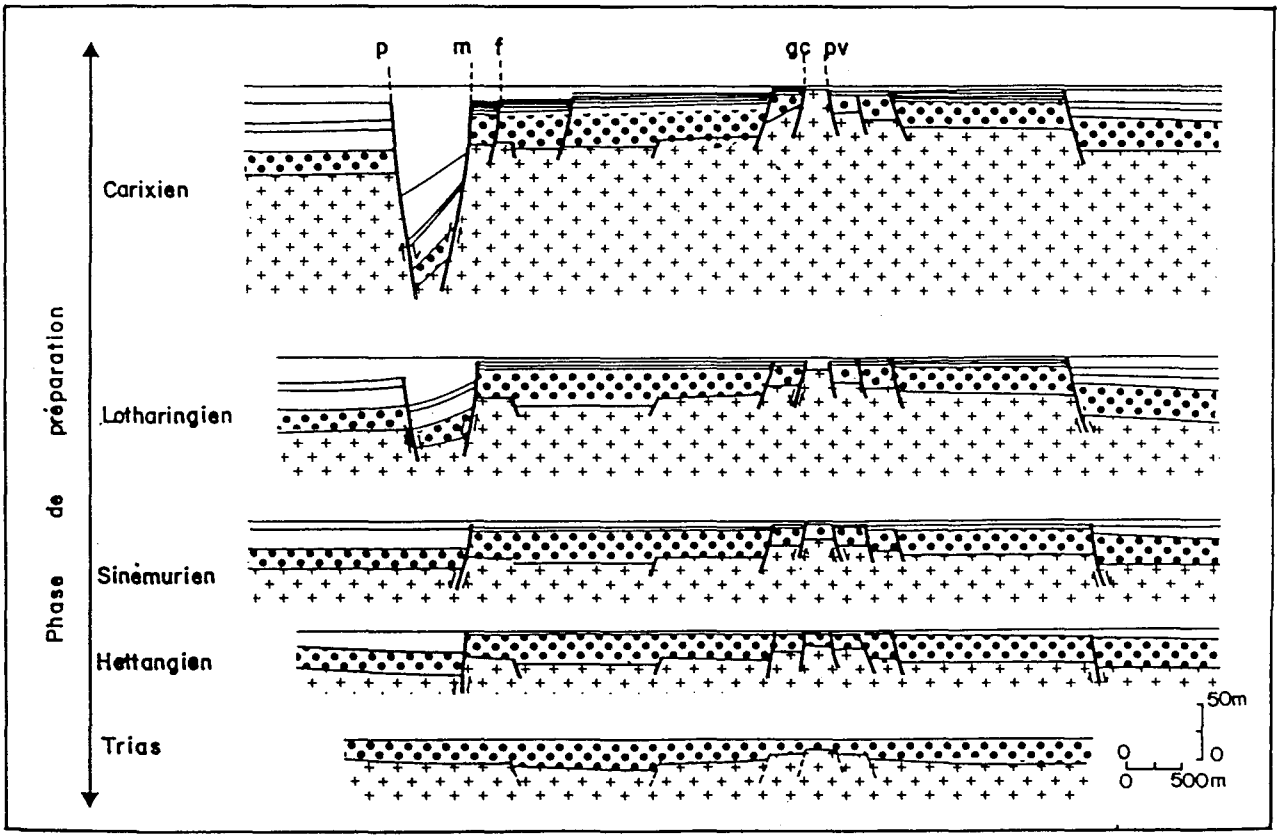


FIG. 9. — Terrains antérift, synrift et postrift entre les Grandes Rousses et le bassin de La Grave au cours du Dogger. La disposition actuelle des affleurements et l'intensité de la déformation alpine ne permettent pas de reconstituer la forme des paléo-grabens situés entre les Grandes Rousses et le plateau d'Emparis.



Bull. Soc. géol. Fr., 1988, n° 4

Fig.3.7. Timing of Rifting in the Pelvoux Massif (from Trift & Graçiansky, 1988).

stratigraphy of the Dauphinois zone. The reader is directed to the following works for greater detail (Debelmas, 1974; Debelmas & Kerkhove, 1980; Graciansky et al, 1981; Lemoine et al 1981; Barfety & Gidon, 1983; Tricart, 1984; Tricart & Lemoine, 1986; Vernet, 1965).

1) The Hercynian Basement.

Basement rocks are exposed in a chain of massifs from Aar in the north to Esterel in the south. The basement suite of the Pelvoux massif comprises gneisses, micaschists, amphibolites, migmatites and granites representing Caledonian basement (augen gneisses dated at 700Ma; Biju-Duval, 1975) and cover sediments (calcareous greywackes, spilites) that have later suffered metamorphism during the Hercynian orogeny and been intruded by anatexial granites and migmatites. These granites have been radiometrically dated at 350-370Ma in Pelvoux, at 275-320Ma in Combeynot and at 285-295Ma in Argentera (Kerckhove, 1979).

2) The Triassic.

The peneplaned basement is directly overlain by Triassic strata. Considerable stratigraphic variations occur in the Triassic between the different Alpine palaeogeographic domains due to localised rifting.

In the Dauphinois Zone, a thin sequence was deposited on the external basement massifs. These consist of a thin (5-10m thick), brecciated, dolomitic unit ("carneules") that is locally underlain by a whitish-grey quartzitic sand (Kerckhove, 1979; Debelmas, 1980). In the Pelvoux massif, these dolomites are overlain by a purple weathering amygdaloidal spilite (10-15m thick) (Barbier, 1954). To the north of the Pelvoux massif, in the Arvinche region, the Triassic is represented by a great thickness (approx.500m, often much thickened by folding and diapirism) of gypsiferous evaporites (Barbier, 1848).

In the Briançonnais Zone, greater subsidence rates resulted in more open shelf conditions and the deposition of a thick carbonate series (Baud & Megard-Galli, 1975; Debelmas, 1955; Tricart, 1984).

Regionally, the lower–middle Trias was a time of marine transgression when the Dauphinois–Briançonnais boundary marked the transition between substantial subsidence and shallow marine carbonate deposition to the southeast, and slow subsidence and evaporitic marine conditions to the northwest (Ziegler, 1988; Lemoine et al, 1986; Tricart & Lemoine, 1986; Coward & Dietrich, 1989).

3) The Lower Jurassic.

During the Liassic radical changes in the palaeogeography of SE. France occurred. The Briançonnais domain, which had steadily subsided during the Triassic, suffered major uplift and became emergent (Fig.3.6). Meanwhile the Dauphinois domain suffered major extension; basement fault blocks being rotated to form a series of grabens, half–grabens and horsts. This produced a Liassic syn–rift basin–and–shoal sedimentation pattern (Barfety et al, 1979; Barfety et al, 1989; Barfety & Gidon, 1983; Grand, 1988; Tricart & Lemoine, 1986; Lemoine et al, 1986) (Figs.3.7 & 3.8).

At least two phases of extension are recognised in the Liassic. **Hettangian extension** was associated with a eustatic sea–level rise which produced palaeo–depths of 2–300m (Roux et al, 1988) but only minor seafloor topography (Fig.3.6). Sedimentation is characterised by monotonous pelagic limestones in the Dauphiné basin that reach 800m in thickness (Kerckhove, 1979) but thin towards the south and west. In the external basement massifs, alternations of thinly–bedded limestones and shales ("lias calcaires") were deposited. This early extensional phase was controlled by NW–SE trending basement faults rather than the NE–SW lineaments that controlled syn–rift sedimentation in the late Liassic extensional phase (Dumont, 1988).

The Domerian–Toarcian sequence comprises black marly shales and interbedded limestones that becomes dominantly shaley with siliceous and pyritous nodules and calcareous lenses towards the top (or "Lias schisteux"). The sequence is very variable across the Dauphinois zone due to syn–depositional rifting of the basement which created a series of half–graben. Condensed footwall sequences contrast greatly with thick shaley hangingwall wedges (Lemoine,

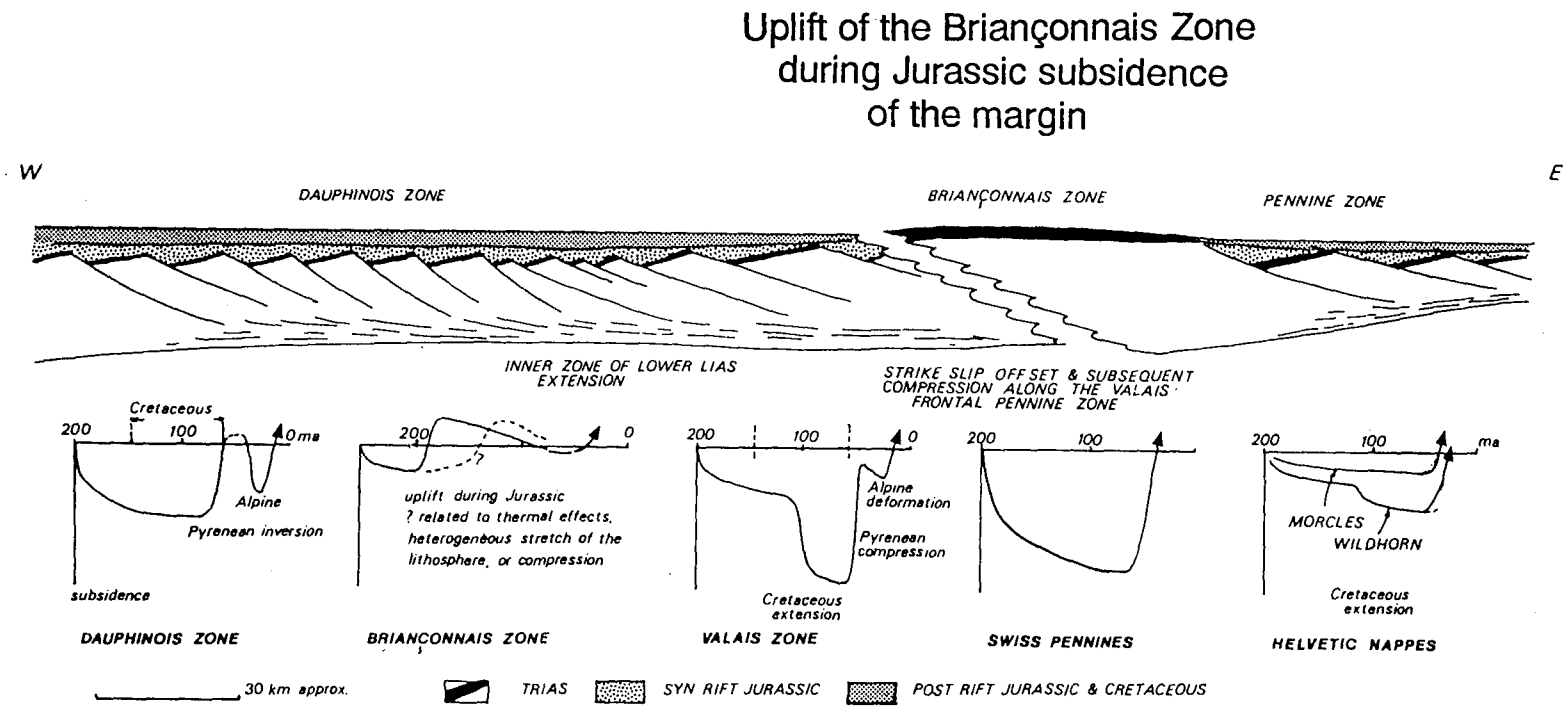


FIG. 11. Simplified cross-section through the western Alps showing the suggested basin structure and subsidence history; from Lemoine *et al.* (1986), Gillcrist *et al.* (1987) and Homewood & Lateltin (1988). The diagrams for the Valais zone, the Swiss Pennines and the Helvetic nappes refer to a section through the central Alps.

Fig.3.8. Sketch-section illustrating the Basin Structure of the Western Alps (from Coward & Dietrich, 1989)

Gidon & Barfety, 1981; Barfety & Gidon, 1983; Grand, 1988). The normal fault displacement varied from hecto- to kilometric scale and olistoliths indicate that the fault planes were topographic features, e.g. Ornon Fault (Barfety et al, 1979).

Major seafloor topography was established during the main Toarcian-Aalenian extensional phase. Approximately 700–1000m of black shales with interbedded limestones accumulated in the Dauphiné basin during this period (Kerckhove, 1979; Rudkiewicz, 1988) but up to 4km of syn-rift sediment are reported in the Bourg d'Oisans half-graben (Coward et al, 1991). The main basin-bounding faults trend NE-SW, being dictated by the Variscan "Cevenoles" trend (Dumont, 1988) (Fig.3.6a).

Halokinesis occurred in association with the Liassic rifting event. The Triassic gypsum of the Dauphinois and Ultradauphinois Zones has flowed to form salt walls and diapirs that rise over 500m above their regional level. Growth primarily affects the upper Lias syn-rift succession. This diapirism has been compared with that of the Atlantic passive margins (Mascle et al, 1988) (Fig.3.9).

4) The Middle and Upper Jurassic.

Following the main late Liassic (Toarcian-Aalenian) rifting phase major seafloor topography was developed and water depths in the Dauphiné Basin locally reached at least 650m (Roux et al, 1988). This deepening was restricted to the Dauphiné basin, however, because platform conditions continued on the Provençal Platform and the Briançonnais zone became emergent at this time. The sedimentary record shows gradually falling relative sea-level between the end of rifting and the onset of foundering in the upper Dogger epoch. The upper shaley Lias is overlain by a Bajocian-Bathonian series characterised by 100–200m of alternating calcareous and shaley marls.

Widespread foundering of the European margin did not occur until Callovian times when oceanic spreading began to the southeast of the Piemont domain. The emergent

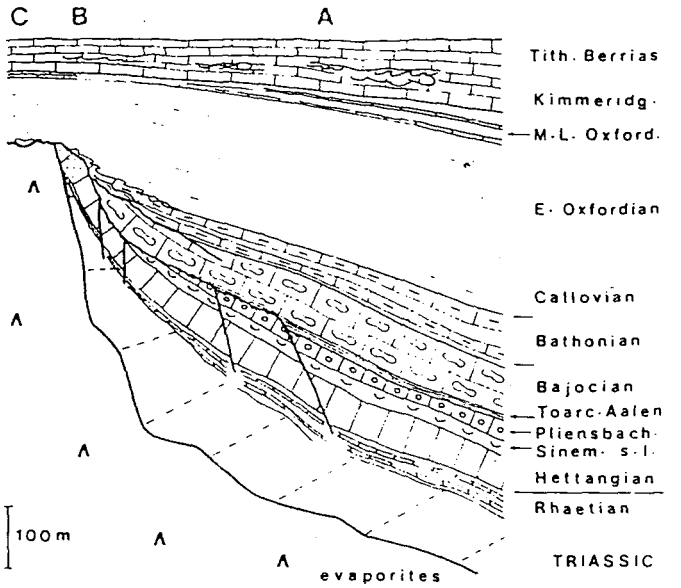


FIG. 5. — Diagrammatic section across a reconstructed salt dome located on the Rouaine sinistral wrench-fault (Alpes maritimes).

[Stratigraphical frame by Dardeau, 1983; unpublished structural data by Dardeau, Delpech, Graciansky and Lemoine].

An Alpine wrench-fault is bounding westward the dome de Barrot and runs SW-NE from Rouaine to Valberg through Daluis in the "Alpes maritimes". It coincidates with a distensional fault and with a line of salt domes of Liassic and Middle Jurassic age reactivated during the Alpine diastrophism. With the difference of the Bonson domes (fig. 4) some of these were active as late as the early Oxfordian and probably reached the sea floor in that time.

The diagram is a composite section built from field data collected at Guillaume (A), Daluis (B) and Valberg (C). The stratigraphical sequence is thinning out when approaching the top of the structure due to numerous gaps and internal unconformities. Synsedimentary faults, slumped beds, conglomerates, and encrustments along bedding surfaces are characteristic of the flanks of the domes.

A rugged topography on the sea floor was probably induced by the salt movement and was preserved for a long time. This could explain (1) the presence of nodular limestones at C instead of well bedded limestone and black shale layers at A in the Middle Oxfordian, or (2) of slumped beds in the latermost Jurassic. The sequence is Barrot type [Dardeau, 1983] widespread in the area (see fig. 4).

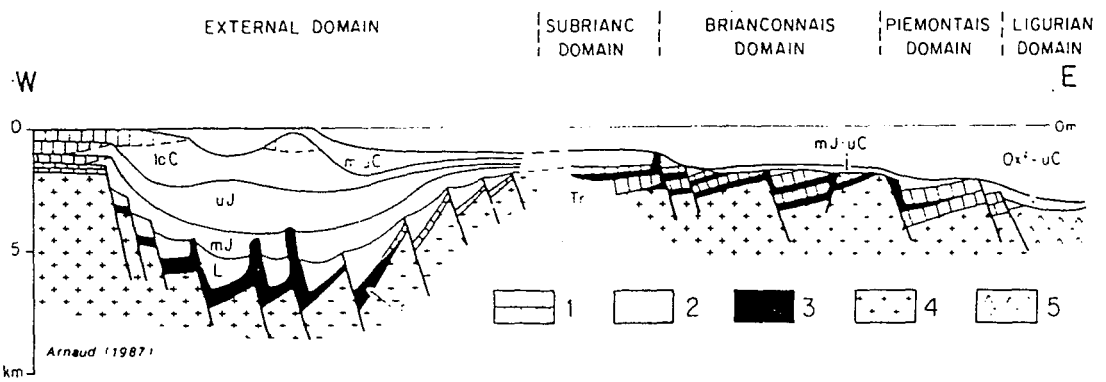


FIG. 10. — Schematic section of the European passive paleo-margin in the western Alps.

The halokinetic structures compare well those of the Atlantic margin (see fig. 8). Halokinesis occurred close to the fractures of the margin in relation with disequilibrium in the sedimentary blanket during rifting.

1 : platform carbonates; 2 : basin formations (clays, marls, marly limestone, limestones, detrital formations); 3 : evaporites; 4 : continental crust; 5 : oceanic crust; L : Lias; Jm : Middle Jurassic; uJ : Upper Jurassic; Oxf. : Oxfordian; loC : Lower Cretaceous; uC : Upper Cretaceous.

Fig.3.9. Halokinesis during Syn-rift Sedimentation on the Tethyan margin of the W.Alps

(from Mascle et al, 1988)

Briançonnais domain subsided dramatically by 2000m (*op.cit.*). In the Dauphiné domain major subsidence continued but water depths did not increase further. This period is represented by the upper Bathonian–Oxfordian "terres noires" sequence of black shales with minor sandy intercalations which includes very characteristic rusty limestone lenses towards the top (Oxfordian). The thickness of these deepwater sediments reaches 3000m (Kerckhove, 1979).

4) The Cretaceous.

During the lower Cretaceous a dominantly shaley pelagic facies continued to be deposited in the Vocontian trough, although shelf deposition persisted to the south on the Provençal platform. Extension related to the rifting of the Bay of Biscay is known (Dumont, 1988). These sediments are not exposed within the Pelvoux–Arvinche region but outcrop to the west in Devoluy, where shaley sedimentation prevailed thoughout the early L.Cretaceous but became increasingly calcareous at the end of the L.Cretaceous. The Senonian limestones lie unconformably upon the lower Cretaceous units indicating the onset of compressional tectonics – the Eoalpine phase. Approximately 1200m of Cretaceous strata were deposited in Devoluy.

5) The Tertiary.

No Palaeocene strata are preserved in this region, and the next sedimentary sequence to be deposited was the U.Eocene–Oligocene "nummulitique" trilogy. This series comprises an uncomformable basal conglomerate that passes into the Nummulitic limestone ("calcaires a Nummulites") marking marine transgression. Deepening of the basin in late Eocene–early Oligocene times is indicated by deposition of the Globigerina marl. Such deepening has also been recorded in the Austrian molasse basin (Nachtmann & Wagner, 1987). Finally, the infilling of the basin by the advancing Alpine flysch is marked by a 400–500m thick series of alternating sands and shales (Gres du Champsaur, Gres des Aiguilles d'Arves). The flysch is marine in the east (S.Pelvoux), but with time, as it prograded westward, the basin became emergent and the flysch is fluvial (Devoluy). This trilogy is a transgressive series that lies unconformably upon all units from Hercynian basement to middle Jurassic cover and represents westward progradation of the Alpine foreland basin.

3.4. STRUCTURAL FRAMEWORK OF THE DAUPHINOIS ZONE.

Much structural data from the western Alps has accumulated during the past 35 years. The regional structure has been reviewed by Trumpy (1960), Debelmas & Lemoine (1970), Ramsay (1963) and recently by Coward & Dietrich (1989). Classically the structure of the Alps has been divided into three zones:-

1) the unmetamorphosed **external Helvetic-Dauphinois Zone** of the European continental margin,

2) the metamorphosed **internal Penninic Zone** of the European continental margin, involving the platform Briançonnais domain and the deepwater Ligurian Piemont domain and

3) the **Austro-Alpine Zone** of the Apulian-Adriatic promontory of the African continent (Ramsay, 1963).

Ophiolite obduction onto the European passive margin during the late Cretaceous heralded the closure of the Ligurian Tethys Ocean and continent-continent collision. Ongoing compression resulted in the overthrusting of the Austro-Alpine Zone onto the Penninic Zone and later, the migration of the deformation front from the internal zone to the external zone in middle-late Oligocene to early Miocene times (Butler et al, 1986; Tricart, 1984; Platt et al, 1989).

The Briançonnais Zone of the Internal Alps is separated from the Dauphinois Zone of the External Alps by the **Frontal Pennine Thrust** (Ramsay, 1963). This simplified definition is complicated by the presence of subzones in the western French Alps (the Sub-Briançonnais and Ultra-Dauphinois Zones). The FPT, which juxtaposes zones of different stratigraphic affinity, is alternatively viewed as a major overthrust zone (Butler et al, 1986) or as a strike-slip zone involving both dextral and sinistral movements (Riccou & Siddans, 1986). The Briançonnais and Dauphinois Zones both represented platform regions during Mesozoic times but, towards the north, they are separated by the deepwater marine realm of the Valais Zone (Trumpy, 1960; Ramsay, 1963).

The Dauphinois Zone comprises the external crystalline massifs and the Subalpine chains. Both thin-skinned (Boyer & Elliott, 1982; Butler, 1983; Butler et al, 1986) and thick-skinned (Beach, 1981; Lemoine et al, 1986; Gillcrist, 1988) models have been applied to the external Alps - distinction between the models hinging upon the interpretation of the role of basement during Alpine deformation. Up to 30km shortening is accommodated by the Subalpine chains (Gratier, Ménard & Arpin, 1989) whilst a further 70km is accounted for in the Ultra-dauphinois Zone (Beach, 1981). This compares with a total restored shortening in the footwall to the FPT in the northern Belledonne massif of 140km (Butler et al, 1986).

Much of the ambiguity is a product of erosion level. The structural style exhibited in strongly uplifted areas where the basement and shoal-and-basin synrift cover is exposed, e.g. N.Pelvoux, is quite distinct from those less-uplifted areas where the post-rift succession is preserved, e.g. the Arvinche and Digne regions. In the former, the non-"layer-cake" stratigraphy and pre-existing structural weaknesses favour thick-skinned inversion geometries. In the latter, the sediment drape is deformed by staircase trajectory, thin-skinned thrusts. **It is important to realise that both structural styles may co-exist and simply reflect the stratigraphic horizon observed.** Structural style is largely controlled by the stratigraphy of the Mesozoic European margin.

In addition, **different structural styles are also associated with successive Alpine deformation phases.** The presence of favourably oriented Mesozoic normal faults in the Alpine foreland meant that at an early stage in Alpine compression these faults were reactivated. For some time, the role of Mesozoic extension and its influence upon the style of the pre-orogenic stratigraphic template used in balanced cross-section construction was largely ignored (Butler, 1983; Butler et al, 1986). This was due to the choice of field study areas in which whole-sale overprinting by later Alpine phases had concealed earlier events. However, studies of the Pelvoux massif (Beach, 1981; Gillcrist et al, 1987; Gillcrist, 1988; Trudgill, 1991), demonstrate that **the Alpine deformation phase was strongly constrained by pre-existing basement**

weaknesses and the form of the passive margin sedimentary sequences.

3.5. PHASES OF ALPINE DEFORMATION.

3.5.1. The Pre-Senonian Phase.

In Devoluy, Upper Senonian limestones unconformably overlie deformed Cenomanian and older strata (Flandrin, 1966; Debelmas, 1970; Siddans, 1979). The deformation comprises a set of E-W trending, N-vergent folds which are assumed to décolle at depth onto a detachment within the cover (Siddans, 1979). Since no thrusts are observed at the surface, it is possible that the folds result from thick-skinned, basement reactivation.

In Pelvoux, deformed Jurassic and older strata are truncated by the late Eocene Nummulitic unconformity. This deformation was interpreted as the deep level expression of the Pre-Senonian phase (Gidon, 1979). This interpretation implies basement involvement in Pre-Senonian deformation. However, the Pre-Nummulitic deformation of Pelvoux is characterised by SW-vergent thrusting and folding of the basement to Jurassic succession (Gillcrist, 1988). It is regarded, therefore, as a distinct and later phase than the Pre-Senonian phase of Devoluy.

The extension of this phase to northern Pelvoux and the Arvinche has been proposed (Barbier, 1963; Barbier & Barfety, 1972). In this area, Nummulitic strata unconformably drape major E-W trending, N-vergent folds in Triassic-Jurassic sediments (Barbier & Barfety, 1972). These anticlines may have initiated due to salt flowage during Tethyan extension (Mascle et al, 1988; Samec et al, 1988) . However, N-vergent folding within the post-rift sequence must result from early Alpine compression (see Chapter 4). Unfortunately the age of these structures can not be tightly constrained and they are described as Pre-Nummulitic although they may belong to the Pre-Senonian (Barbier & Barfety, 1972), Pyrenean-Provencal (Butler et al, 1986) or Pre-Nummulitic phase. The orientation of the folds would tend to exclude the latter, but due to the variations in stratigraphy between Barronies, Diois, Devoluy and the Arvinche, distinction

between the pre-Senonian and Pyrenean-Provencal phases can not be made on the basis of structural style.

In addition, N-vergent thrusting has been observed on the northern margin of the Pelvoux massif (*op.cit.*) and towards the west in the Grandes Rousses massif (Barfety & Gidon, 1983; Desthieux & Vernet, 1970) and southern Belledonne massif (Barfety & Gidon, 1990). This deformation is also thought to be Pre-Senonian (Barbier, 1963; Barbier & Barfety, 1972).

3.5.2. The Pyrénéan-Provençal Phase.

This phase, which was referred to as the "plis anciens" (Goguel, 1963), is best developed in the Alpes Maritimes. E-W trending folds belonging to this phase have been described in Baronnies (Flandrin, 1966), Dignes (Graham, 1978) and southern Devoluy (Siddans, 1979). The folding affects the Mesozoic cover sediments and is locally overlain by Eocene-Miocene flysch deposits. Geophysical evidence suggests that basement is not involved in these folds (Flandrin & Weber, 1966) but more recent stratigraphical studies show a marked coincidence of Mesozoic extensional geometries and later fold trends (Dardeau, 1988).

The northward continuation of this phase is unknown. The E-W trending folds of the Arvinche may be of this age (Butler et al, 1986).

3.5.3. The Pre-Nummulitic Phase.

In Pelvoux and the Arvinche, the subdivision of Alpine deformation *sensu stricto* into Pre- and Post-Nummulitic phases has been made possible by the deposition of an unconformable syn-orogenic flysch series of late Eocene-Oligocene age (Debelmas, 1980). Lory (1894) noted the discordance of these flysch sediments upon all strata from basement gneiss to Jurassic shales in southern Pelvoux. He concluded that, in SW Pelvoux, pre-Nummulitic deformation was much more intense than later movements and wondered at the regional extent of this early deformation; a view shared by Gidon (1979).

Pre-Nummulitic deformation in both southern and eastern Pelvoux has been described in terms of NW-SE directed compression (Gidon, 1979; Bravard & Gidon, 1979) (Fig.3.10) and related to the Pre-Senonian phase of Devoluy (Gidon, 1979). Kinematic analysis of fault planes has led to the re-interpretation of this deformation in terms of WSW- vergent basement thrusting (Gillcrist, 1988). The data presented in Chapter 4 supports the latter model.

This deformation produced significant uplift (1km minimum) that brought basement to the surface prior to the deposition of the Nummulitic flysch series, and it remains anomalously above its regional level in this area today. In southern Pelvoux, the pre-Nummulitic structures are characterised by "pli-faillé" or fault-propagation folds (Fig.3.11). These structures develop as folds that are breached at a late stage by low angle thrusts (Debelmas et al, 1989, Gidon, 1989). The extent of Alpine inversion is unknown. However, in northern Pelvoux the reactivation of Mesozoic normal faults has been proposed (Beach, 1981; Gillcrist et al, 1987; Coward et al, 1991) (Fig.3.12).

To the north of Pelvoux, Barbier (1956, 1963) recognised Pre-Nummulitic deformation in the substratum to the Aiguilles d'Arves flysch. He proposed that the main period of "Arvinche" uplift and imbrication was of Palaeocene age and restricted to the Pelvoux and Arvinche regions. Beach (1981a&b) argued that these structures were Post-Nummulitic because the Nummulitic unconformity has been sheared. The fact that the unconformity has been sheared during Post-Nummulitic deformation does not preclude Pre-Nummulitic deformation however. In fact, in order to maintain lateral continuity with Pre-Nummulitic structures in the Combeynot massif, the presence of Pre-Nummulitic thrusting in the Arvinche is probable (Trudgill, 1991).

Barbier & Barfety (1972) described E-W trending folds in the hangingwall of the Ultra-Dauphinois Thrust as pre-Nummulitic, a fact supported by field observations of the base of the Aiguilles d'Arves flysch during this study. These structures are, however, quite distinct from the WSW-vergent Pre-Nummulitic structures of Pelvoux and seem more closely related to

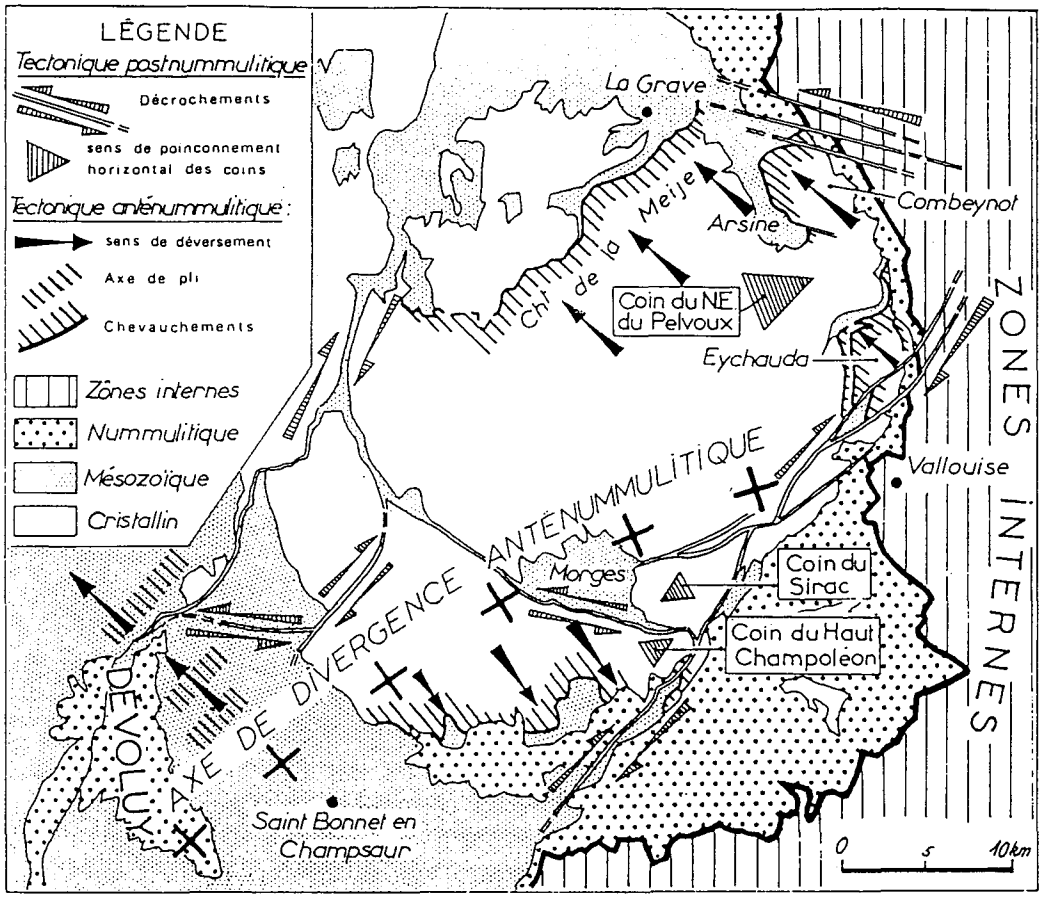


Fig.3.10. Superposition of Alpine Phases in Pelvoux (from Gidon, 1979).

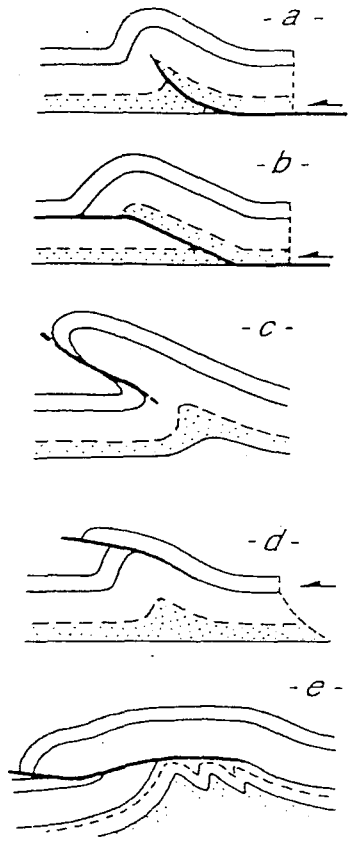


Fig.11. - **Schémas théoriques**, montrant les differences entre 5 cas de relations qui peuvent etre envisagées entre plis et chevauchement.

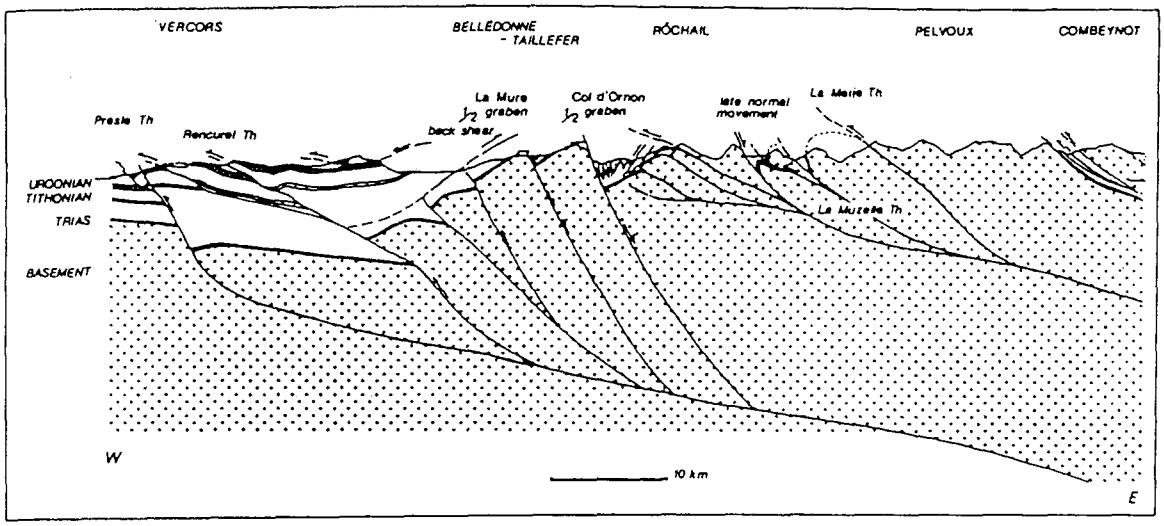
a - pli d'amortissement ; b - pli de rampe ; c - pli-faille ; d - chevauchement par décollement dans le flanc long d'un pli (processus examiné dans cette note) ; e - chevauchement par rupture d'un banc "d'armature" *(Strut thrust*, selon Willis *et al.* 1934). Dans ce dernier schema l'apparition du chevauchement etait attribuée à la compétence plus forte des couches de la voûte vis à vis de celles du coeur du pli, replissées en accordéon. Or dans aucun des cas présentés ici un tel facteur ne peut être mis en cause car la rupture se produit au sein meme de formations relativement homogènes et non pas à leur limite.

Fig.11. - ***Five kinds of possible interrelations between a thrust and an anticline*** : *a - the fault-propagation fold ; b - the ramp fold ; c - the "pli-faille" (stretch-thrust) ; d - the kind of thrusting here described ; e - the strut-thrust (after Willis et al. 1934)*

Fig.3.11. Fold/Thrust Relationships

Types (a) & (c) are commonly developed in the Pre-Nummulitic structures of southern Pelvoux.

(from Gidon, 1989)



— **Cross section from the Vercors, through the southern [illegible]nne, Rochail and Pelvoux basement massifs, to eastern Pelvoux (line shown on Figure 6).**

It shows the structure of the early extensional faults, where they are still well preserved, the principal thrusts and late normal faults. The positions of the Isere and basal Belledonne faults are determined by geophysical techniques (see Menard, 1979; in press). Some of the thrust sheets have moved in a SW direction, out of the plane of section.

FIG. 10. — *Coupe à travers le Vercors, l'extrémité sud de Belledonne, le massif du Rochail et celui du Pelvoux (ligne de coupe marquée sur la Fig. 6).*

On voit la structure des failles normales tardives, des chevauchements alpins et des failles normales antérieures, là où elles sont bien préservées. Les positions de la faille de l'Isère et du chevauchement sous Belledonne sont déterminées d'après la géophysique (voir Ménard 1979, Ménard et Thouvenot, sous presse). Certaines des structures chevauchantes se sont déplacées vers le S.W., hors du plan de coupe.

Fig.3.12. Inversion of Mesozoic half-graben in northern Pelvoux (from Gillcrist et al, 1987)

the E–W trending Pre–Senonian phase of Devoluy (Barbier, 1963; Barbier & Barfety, 1972) or Pyrénéan–Provençal phase of the Alpes Maritimes (Butler et al, 1986).

3.5.4. The Post–Nummulitic Phase.

The Main Alpine or Post–Nummulitic phase (Siddans, 1979) established a through–going breaching system of low angle thrusts. The Frontal Pennine Thrust (FPT), the Ultradauphinois Thrust (UDT), the Belledonne Thrust, the La Meije–Sirac Thrust, the Devoluy Thrust and the Digne Thrust were all active at this time.

The FPT carried the Briançonnais domain across the Dauphinois domain by in excess of 45km on the evidence of the Pre–Alpine klippen (Niggli, 1978). The coincidence of the FPT with a major Mesozoic palaeogeographic boundary suggests that the FPT was strongly constrained by inherited extensional structures (Ramsay, 1963). In the footwall of the FPT, the Ultradauphinois Zone was carried across the Dauphinois Zone on the UDT which exploits the Triassic evaporite sequence as an easy–slip horizon (Barbier, 1948, 1956). This WNW–vergent thrust breaches the Nummulitic unconformity. Towards the south, the evaporites diminish and a number of thrust splays replace the UDT. These thrusts were regarded as Pre–Nummulitic by Barbier (1956, 1963), but Beach (1981) argued that they were Post–Nummulitic since they are associated with the UDT. This interpretation introduces lateral continuity difficulties with the Pre–Nummulitic structures of Combeynot (Trudgill, 1991).

Studies of the Pelvoux massif conclude that Post–Nummulitic deformation is subordinate to Pre–Nummulitic deformation within the massif (Lory, 1894; Gidon, 1979; Bravard & Gidon, 1979). The La Meije–Olan–Sirac Thrust system represents a major basement breaching thrust system with displacement in excess of 12km (Gillcrist et al, 1987; Gillcrist, 1988). The Post–Nummulitic deformation phase is best observed within the Nummulitic flysch series (Tricart, 1980; Plotto, 1977). The basal Nummulitic limestone member is often relatively undeformed because the majority of the displacement is accommodated by a widespread detachment within the overlying Globigerina marl member. Shearing of the overlying Gres du

Champsaur flysch (Plotto, 1977) produces marvellously-exposed fold trains typical of mechanical multilayer rheologies (Price & Cosgrove, 1990) and layer-parallel shears accompanied by minor thrust ramping (Gayer, *pers.comm.*). Post-Nummulitic shearing is consistently WSW-vergent in SE. Pelvoux, but local variations occur in Champsaur.

The Devoluy Thrust carries Upper Jurassic to Cretaceous strata over Oligocene-age flysch deposits and is associated with a NNW-SSE trending, W-vergent fold set. These structures are superimposed upon the E-W trending pre-Senonian deformation (Flandrin, 1966; Siddans, 1979). The resultant field outcrop pattern exposed in the Jurassic and lower Cretaceous rocks is a complex "Type 2" fold interference pattern (Ramsay, 1967; Ramsay & Huber, 1987). The thrust can be traced southward towards the Dome de Remellon and Digne.

In Pelvoux, the Nummulitic unconformity was uplifted by at least 2km during Post-Nummulitic deformation since, on the southern margin of Pelvoux, the tilted unconformity descends from the peak of the Aiguille de Cedera to the Champoleon valley floor (Fig.3.13) and, in eastern Pelvoux, even greater vertical exposure occurs at Pointe de Clapouse.

Post-Nummulitic transport directions vary between WNW-vergent in the NW external Alps (Butler et al, 1986) and WSW- or SW-vergent in the SW external Alps (Platt et al, 1989; Fry, 1989) (Fig.3.14). Possible mechanisms responsible for this fanning are presented by Coward et al (1991) (Fig.3.15), including rotation models (Heller et al, 1989; Vialon et al, 1988; Gratier et al, 1988), lateral escape (wedging) and gravitational collapse models (Platt et al, 1989). Fry (1989) argues that two recurrent transport directions exist in the southwestern French Alps, one WNW-vergent and the other SW-vergent. Platt et al (1989) suggest that the transport directions form a divergent fan. The divergence may be a product of Mesozoic extensional inheritance (Gillcrist, 1988).

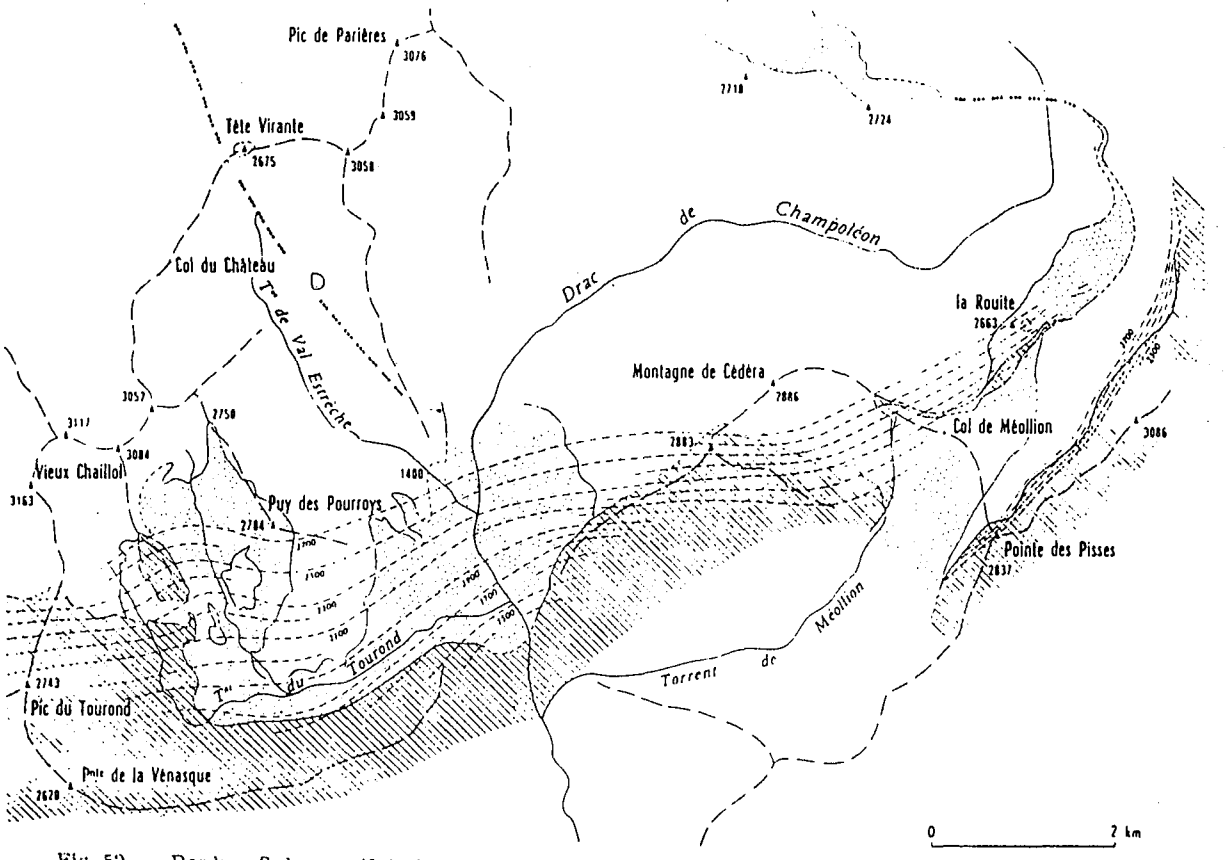


Fig. 52. — Bordure S du massif du Pelvoux. Position de la base du Priabonien (pour la légende, voir fig. 7).

Fig.3.13. Tilting of the Nummulitic Unconformity : Evidence of Pre-Nummulitic Uplift (from Vernet, 1965).

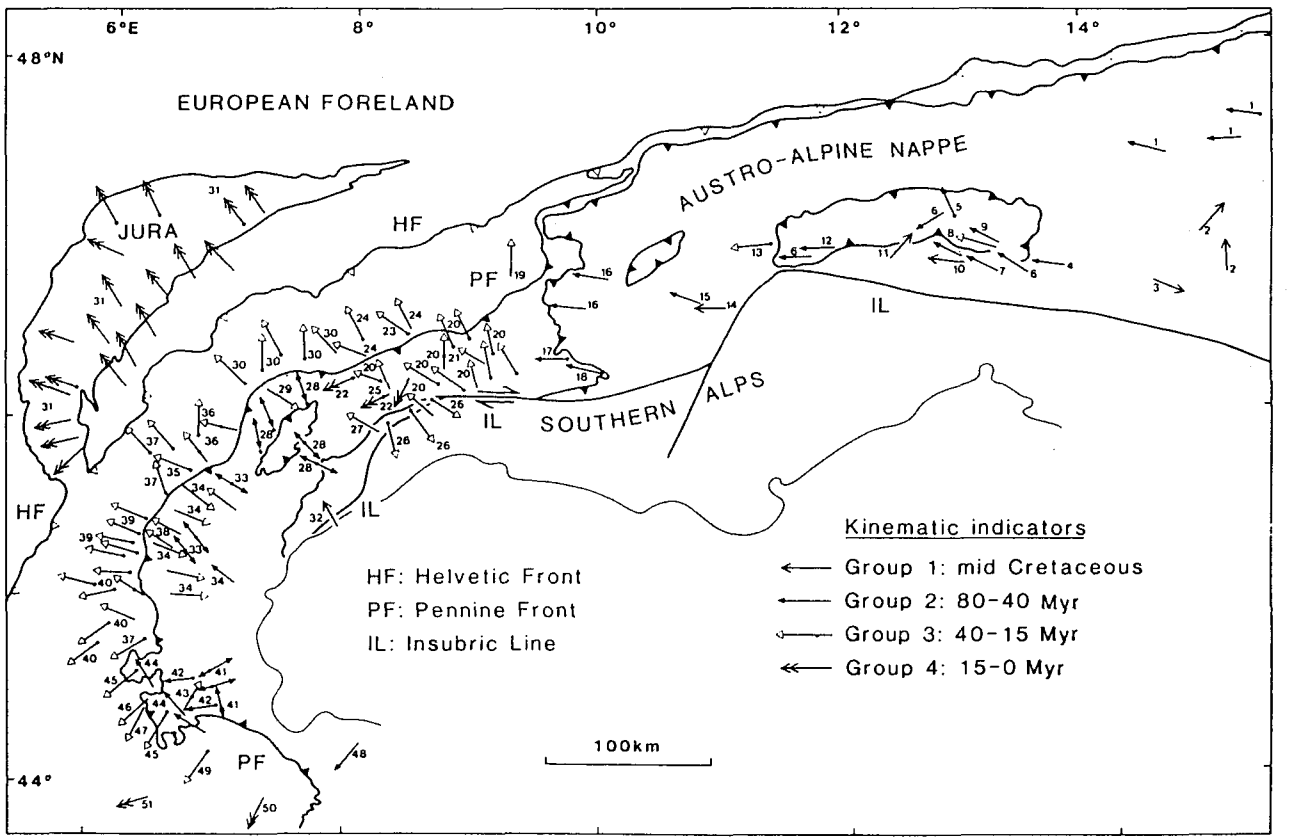


Fig. 1 Kinematic data from the Alps plotted by age. Arrow shows sense of shear. Double-headed arrows indicate that the sense is not known. The heads of arrows lie at the location of data, unless indicated by a filled circle on the arrow shaft. Note that many locations show kinematic indicators of two or more ages: these have been separated slightly on the map for clarity. For the Jura we only show data related to folding[42]. Map projection is conical orthographic.

Fig.3.14. Thrust Kinematic Data from the Western Alps (from Piatt et al, 1989).

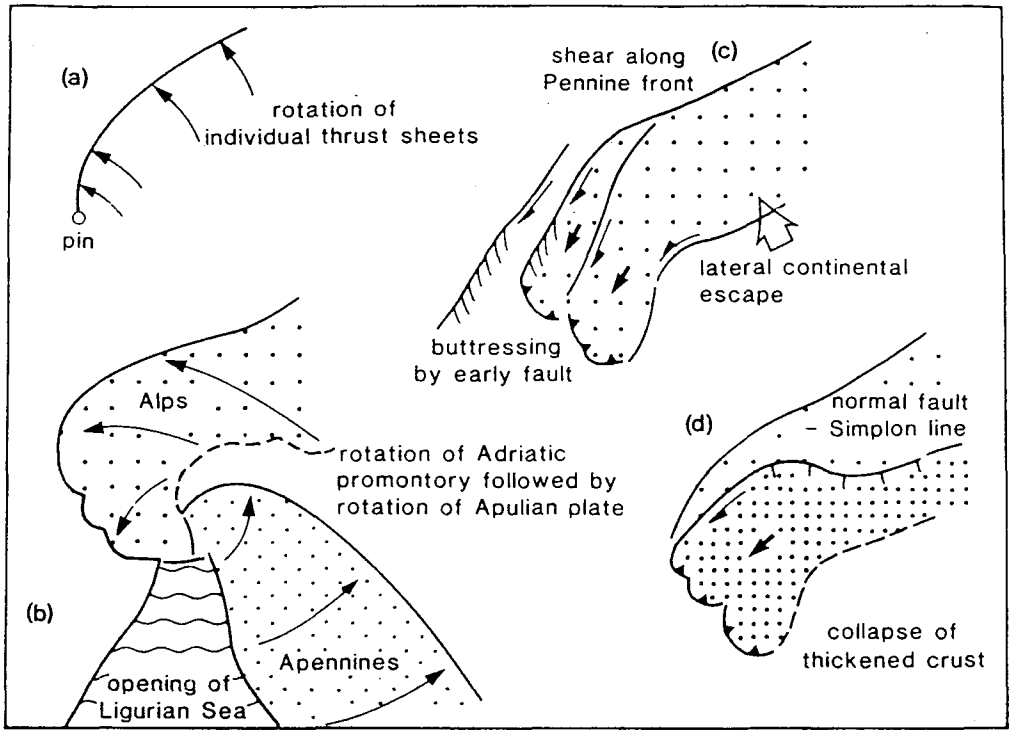


Fig.3.15. Possible Mechanisms for Thrust Transport Direction Variations in the W.Alps (from Coward et al, 1991).

CHAPTER 4.

FIELD DATA FROM THE WESTERN FRENCH ALPS.

The aim of the Alpine part of this research project has been to examine the superposition of successive compressive phases during the Alpine orogeny. The superposition of these phases upon each other and the pre-existing Mesozoic extensional structures has produced a very complex outcrop pattern. Four distinct phases of Alpine deformation have been documented in the external zones of the French Alps; the pre-Senonian, Pyrenean-Provencal, pre-Nummulitic and post-Nummulitic phases. Three of these phases have been recognised in the Pelvoux region but most of the fieldwork described below has focussed upon distinction between the pre- and post-Nummulitic phases.

This distinction can be made where the base of the Nummulitic flysch series outcrops. These localities lie in a belt on the southern, eastern and northern margins of the Pelvoux massif (Fig.4.1). **The field data is presented in this chapter under the following geographical subdivisions :-**

1) Southern Pelvoux; comprising the St. Bonnet, Muande, Tourond and Cedera outcrops (section 4.1).

2) Southeastern Pelvoux; comprising the Meollion, Rougnou, Selle, Fournel and Dourmillouse outcrops (section 4.2).

3) Eastern Pelvoux; comprising the Chambran-l'Yret and Ailefroide outcrops (section 4.3).

4) the Arvinche region, north of Pelvoux (section 4.4).

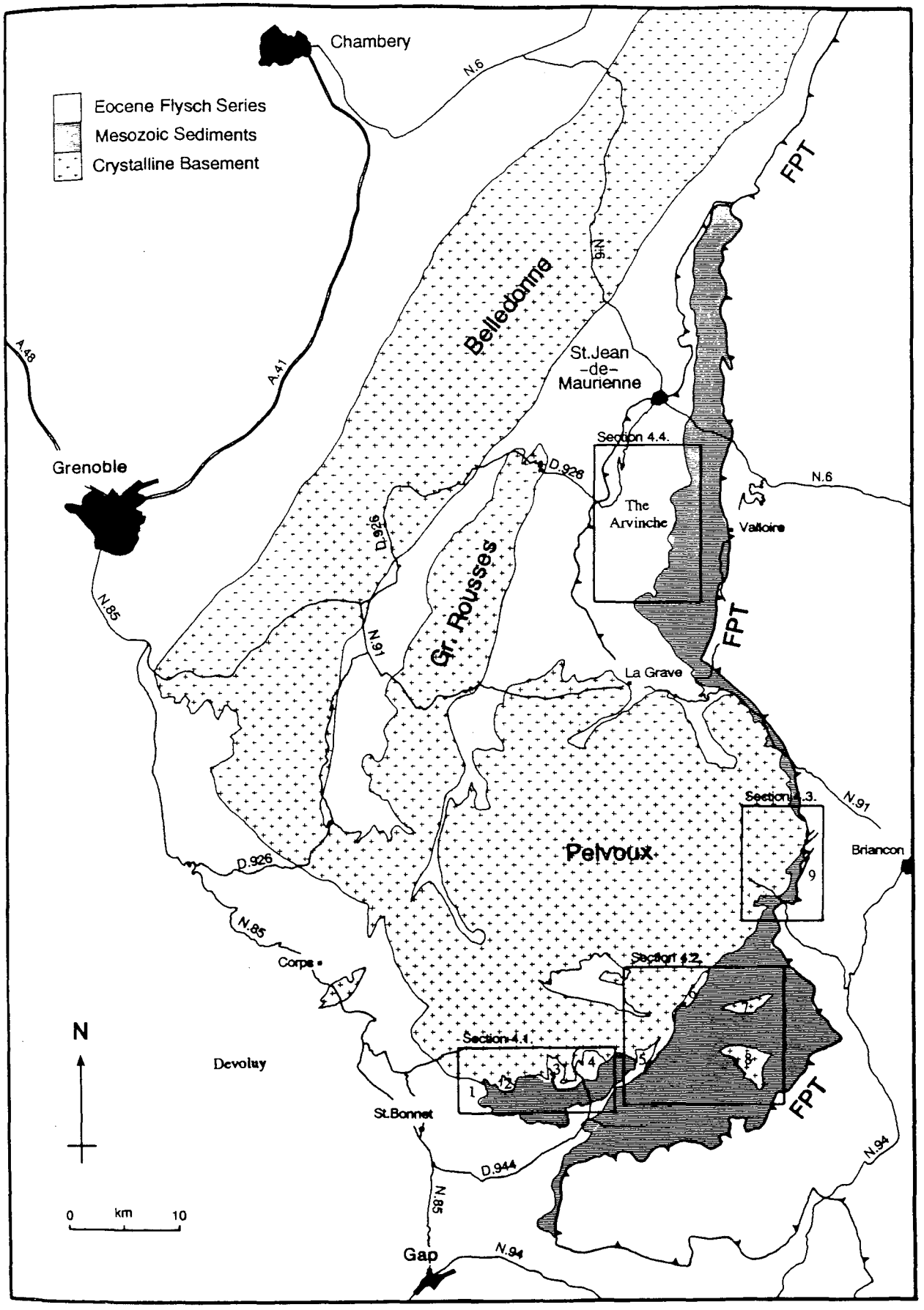


Fig.4.1 MAP OF THE PELVOUX BASEMENT MASSIF,

showing the location of the principal study areas (boxed areas) and main stratigraphic units of the Dauphinois Zone

1, St.Bonnet; 2, La Muande; 3, Tourond; 4, Cedera; 5, Meollion-Rougnoux; 6, La Selle; 7, Fournel; 8, Dourmillouse; 9, La Blanche-Chambran

4.1. FIELD DATA FROM SOUTHERN PELVOUX.

Southern Pelvoux has long been cited as a region characterised by the superposition of two deformation phases (Lory 1894). The uplift and erosion history of this area is such as to expose the basement and Mesozoic sequences and yet preserve the Nummulitic flysch series; permitting distinction of pre- and post-Nummulitic phases. In most of the Dauphinois zone however, erosion has removed the flysch series and distinction of these phases is more subtle. On the basis of the exposures in southern Pelvoux the presence of an early Alpine deformation phase is indisputable. Its original extent is, however, unknown.

The deformation phases are most clearly distinguished on the basis of truncating or breaching relationships between the structures and the Nummulitic unconformity. Tectonic transport direction may also be used diagnostically, but due to variations within a single deformation phase can only be used in conjunction with the criterion of structural style. The pre-Nummulitic deformation is characterised by thrusts that carry basement slices over large footwall synclines of basement and Mesozoic series rocks ("pli-faille" in the French literature). The post-Nummulitic deformation occurred at higher structural levels and strain rates, being charaterised by brittle deformation of the basement and Mesozoic series and layer-parallel shearing and folding of the flysch series. Mesozoic extensional geometries have strongly constrained both Alpine phases.

The pre-Nummulitic phases resulted in the uplift of central Pelvoux, exposing basement suite rocks at the surface. Towards the south the degree of uplift lessened so that a regional southward tilt permitted the preservation of Mesozoic strata on the southern margin of the massif prior to flysch deposition. During post-Nummulitic deformation the Pelvoux massif was uplifted still further to cause the present-day southward tilt of the Nummulitic unconformity of approximately 30° (Vernet,1965).

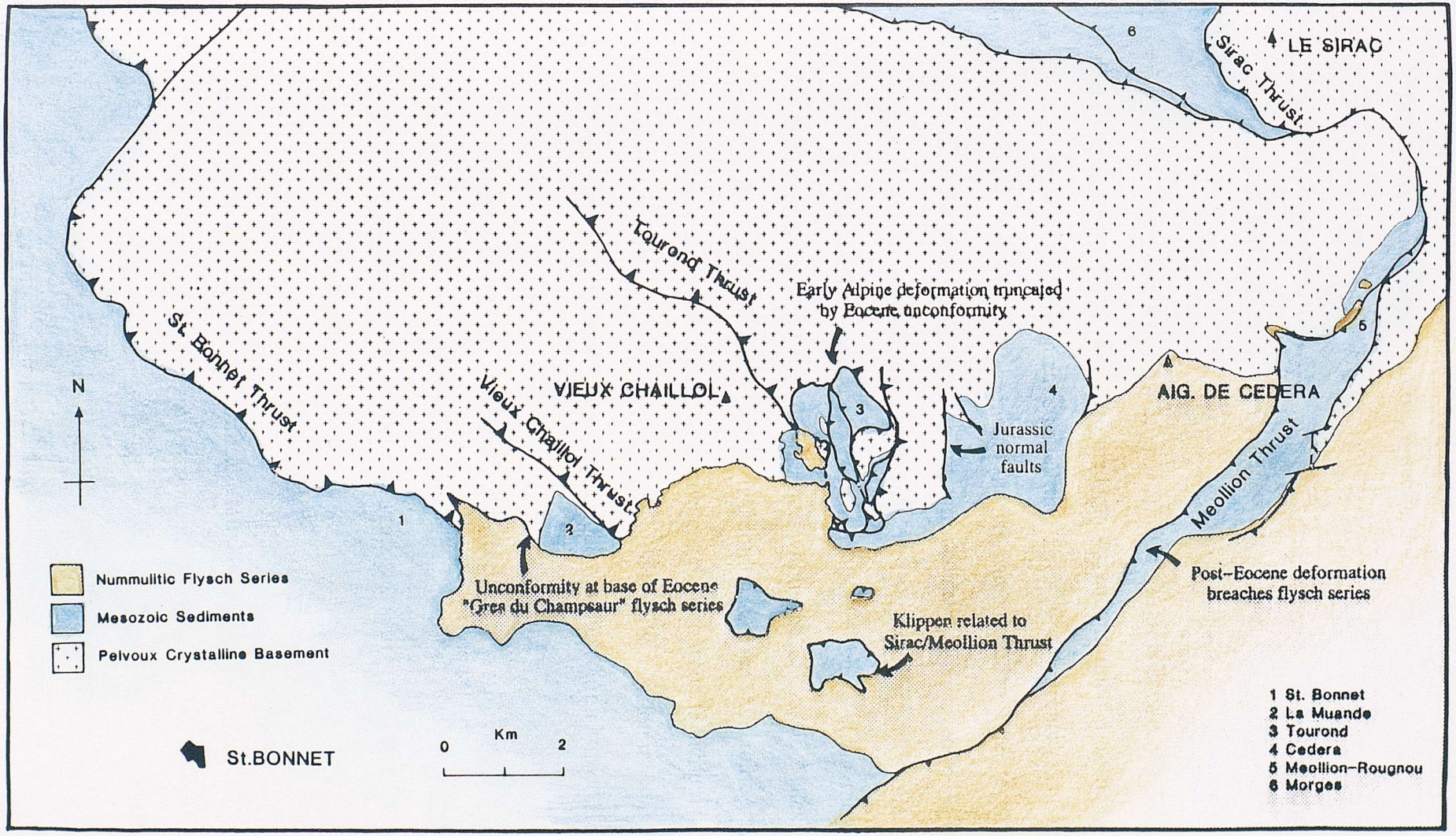


Fig.4.2 MAP OF THE CHAMPSAUR REGION, SOUTHERN PELVOUX.

Three localities are described in this section :-

1) **St. Bonnet** (section 4.1.1),

2) **La Muande** (section 4.1.2),

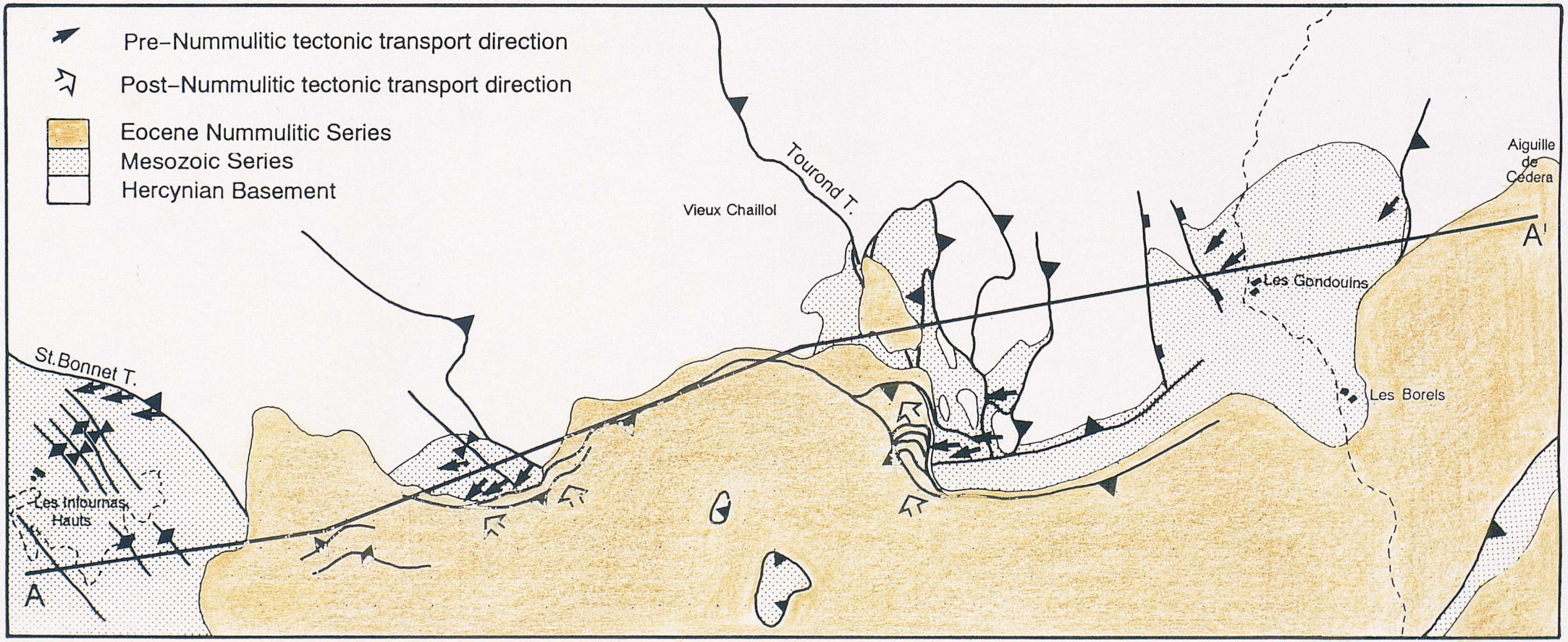
3) **Tourond and Cedera** (section 4.1.3).

At each locality (Fig.4.2) the Triassic-L.Jurassic cover sediments have been overthrust by basement slices. This deformed pile is unconformably overlain by the Eocene Nummulitic series which in this area is called the "Gres du Champsaur". The pre-Nummulitic deformation phase was WSW-SW vergent and is characterised by the development of large synclines in the footwall to major basement thrust slices (Figs.4.3 & 4.4). These structures develop initially as tipfolds to thrusts within the basement that are subsequently breached as displacement on the thrust propagated towards the surface. On reaching the overlying cover sequence the thrust tends to flatten and propagate within the L.Jurassic shales, e.g. St.Bonnet, Vieux Chaillol and Cedera thrusts (Fig.4.4). In addition to such "pli-faille", the pre-existing Tethyan extensional faults were locally reactivated, e.g. Tourond/Gondouins thrust (Fig.4.4). Although a net extensional displacement is still recorded on the Gondouins fault this is interpreted as an artifact of the "ile flotante" breaching geometry of the Tourond thrust.

Hence **the main basement architecture of southern Pelvoux is of pre-Nummulitic age**. A high-level NW-vergent deformation that affected the Eocene flysch series but caused only slight shearing of the Mesozoic strata was superimposed upon these early structures in post-Nummulitic times.

The basement thrusts of St. Bonnet, Vieux Chaillol and Tourond ramp laterally towards the SSE. The thrust planes rise from the basement to the middle Jurassic shales ("terres noires"). The ramp delineated the margin of the Pelvoux massif but was peneplaned prior to the deposition of the Nummulitic series. A slight remnant topography at the time of Nummulitic transgression has been inferred from thickness variations within the Nummulitic limestone

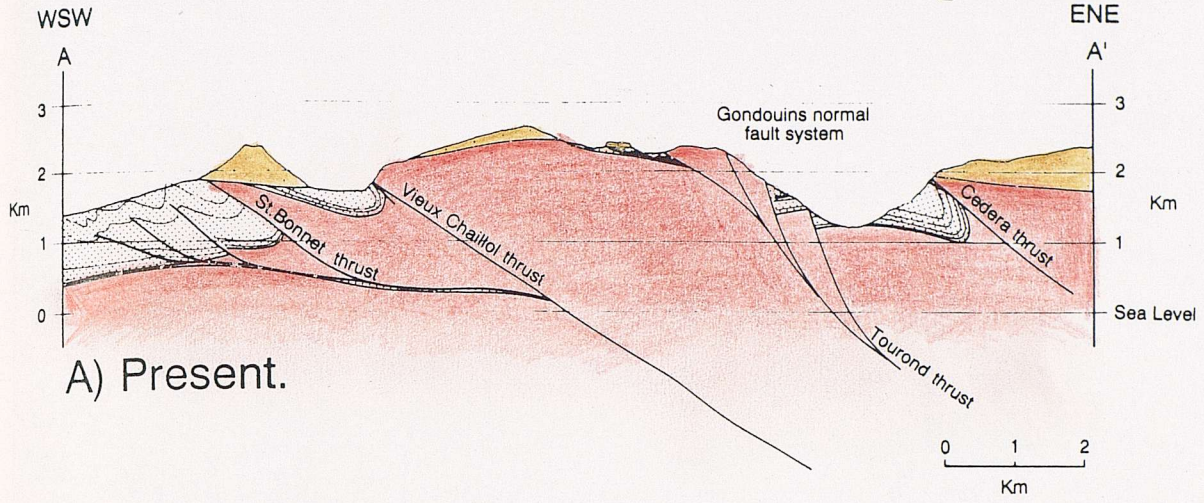
Fig.4.3 MAP OF SOUTHERN PELVOUX.



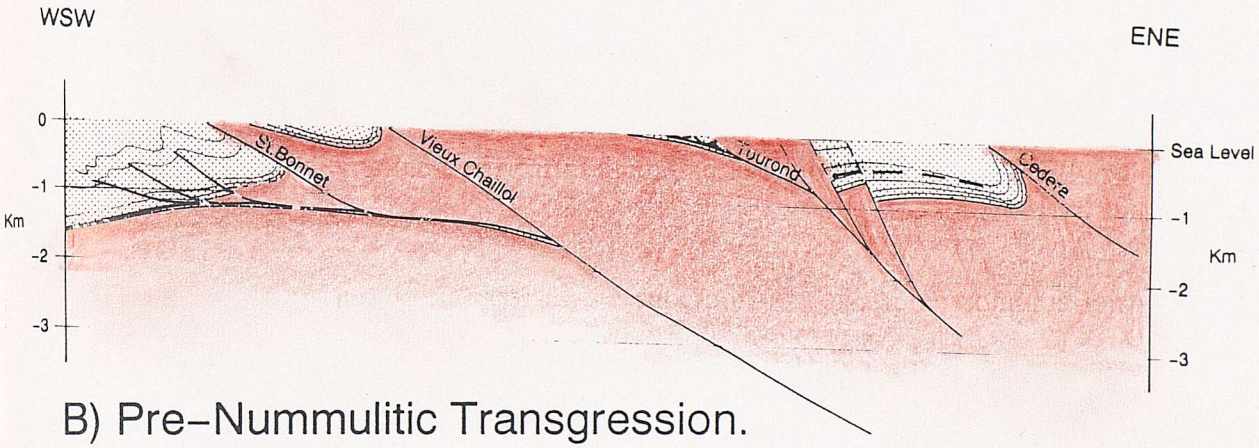
The main structure of southern Pelvoux is illustrated in Section A–A' which is constructed parallel to the pre–Nummulitic transport direction. The post–Nummulitic phase resulted in deformation normal to this direction.

Section A–A' is shown in Figure 4.4

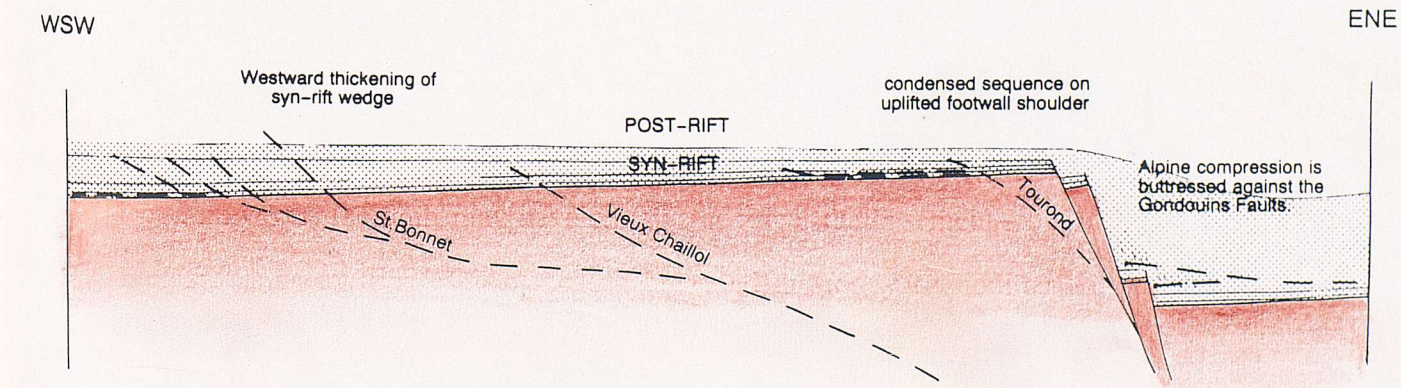
Fig.4.4 PARTIALLY RESTORED CROSS-SECTIONS ALONG LINE A-A'. For location see Figure 4.3.



A) Present.

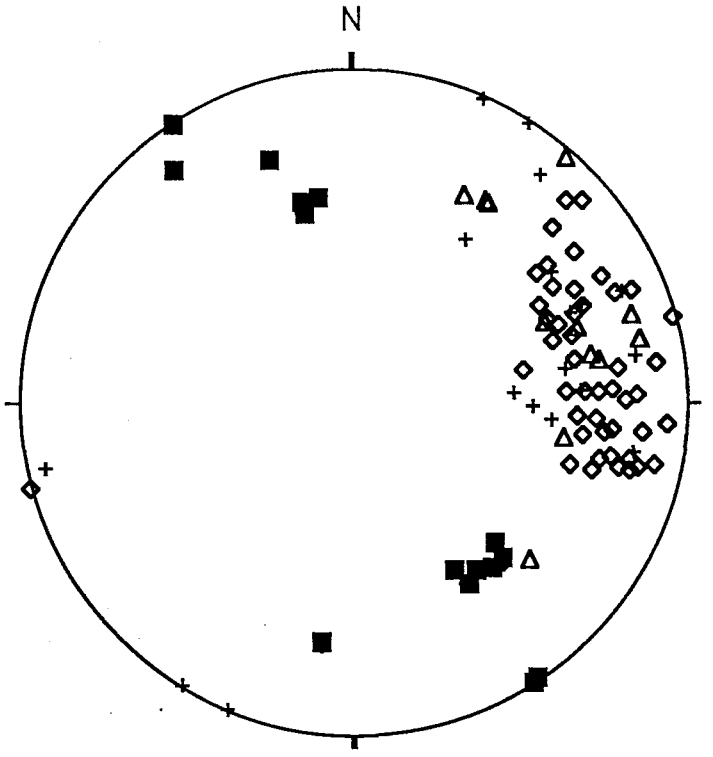


B) Pre-Nummulitic Transgression.



C) Pre-Alpine Extensional Structure.

Mineral Lineation data from Southern Pelvoux



Key

Pre-Nummulitic Data

◇ 46 Data from St.Bonnet
△ 12 Data from Muande
\+ 16 Data from Tourond

Post-Nummulit ic Data

■ 13 Data from Tourond

(S.Gupta, *pers.comm.*). Today the margin of the massif lies at the same place indicating that during both the pre- and post-Nummulitic deformation phaes a lateral culmination formed at this point, perhaps reflecting a deep control such as a pre-existing extensional structure.

A feature of the southern Pelvoux section is the absence of Dogger, Malm and Cretaceous rocks in the footwall of the principal basement thrusts. Sedimentological and stratigraphical evidence from the Vercors, Devoluy and Ubaye regions argues that these units were deposited in Pelvoux but have been subsequently removed. **A detachment within the Dogger shales, perhaps the "terres noires" unit, is inferred**. This separated the basement-Liassic sequence from the Dogger-Cretaceous sequence during pre-Nummulitic deformation. Hence, the Dogger-Cretaceous sequence always remained structurally higher than the basement-Liassic sequence and following pre-Nummulitic uplift of approximately 2-3km, the former was eroded and is preserved only at the less-uplifted margins of the massif. The existence of such a decollement throughout Pelvoux would account for the absence of Dogger and younger sediments in the "pinched-in synclines" (Tricart & Lemoine 1986) of Pelvoux.

Figure 4.4. emphasises the superposition of Alpine compression upon inherited Mesozoic extensional structures. To a large extent the Dauphiné half-graben system is unrecognisable, in southern Pelvoux, because the erosion level only preserves the pre-rift and lower syn-rift sediments. The Gondouins fault system, however, had a minimum displacement of between 1500-2000m and controls the location of the Champoléon valley. The faults, which still dislay normal displacement, trend NNW-SSE and dip 60-70°E. Due to their steep dip at the surface they were not reactivated during Alpine compression but acted as buttresses (Gillcrist et al, 1987), producing tight folds with a steep axial planar cleavage in the Liassic half-graben fill. At depth, however, the fault may have been reactivated. Near the basement-cover contact the thrust departed from the existing fault plane and breached the footwall shoulder of the normal fault block, creating an "ile flotante".

4.1.1. St. Bonnet/Les Infournas Hauts.

St. Bonnet is situated 14km north of Gap, in the Drac valley. 4km to the northeast lies the village of Les Infournas Hauts which lies in the shadow of Le Cuchon (2375m) and the Pic Queyrel (2440m) (Fig.4.5.). From Les Infournas Hauts a forest path leads to these summits via the Col de l'Escalier (2167m).

Le Cuchon is capped by Hercynian micaschists that override the Mesozoic cover on the St. Bonnet thrust. The Triassic–Jurassic strata which comprise dolomites, spilites, limestones and marly shales are tightly–folded about SSE–plunging axes. The upper Lias, Dogger and Malm shales possess a strong, N–S trending cleavage foliation **indicating W–vergent compressive strain**. Towards the summit of Pic Queyrel, at Le Barry (2272m), these structures are truncated by the basal conglomerate of the late Eocene–Oligocene Nummulitic Flysch Series. The St. Bonnet thrust and its associated footwall folds are therefore of pre–Nummulitic age.

At the summit of the Pic Queyrel post–Nummulitic deformation is observed within the flysch. Two styles of deformation are observed. Firstly, large NE–SW trending, NW–vergent folds are exposed on the eastern flank of Pic Queyrel. Secondly, the alternating sands and shales have deformed by layer–parallel shearing, like a "deck of cards", producing slight offsets in the sandstone bands and a strong shear foliation inclined at 20° to the bedding in the interbedded shales. This deformation is of a high–level, superficial nature and is not observed in the base–ment or Mesozoic strata.

The St. Bonnet thrust can be traced northwestwards from Le Barry, where it is truncated by the Nummulitic unconformity, across the Severaissette valley to Petit Chaillol. The fault trace then swings northeastwards up the Severaisse valley for 2km before resuming its NW–SE trend.

At outcrop the thrust carries basement rocks over inverted Triassic and Liassic strata. In the southeast, however, thrusted pods of Triassic spilite occur in the footwall of the St. Bonnet

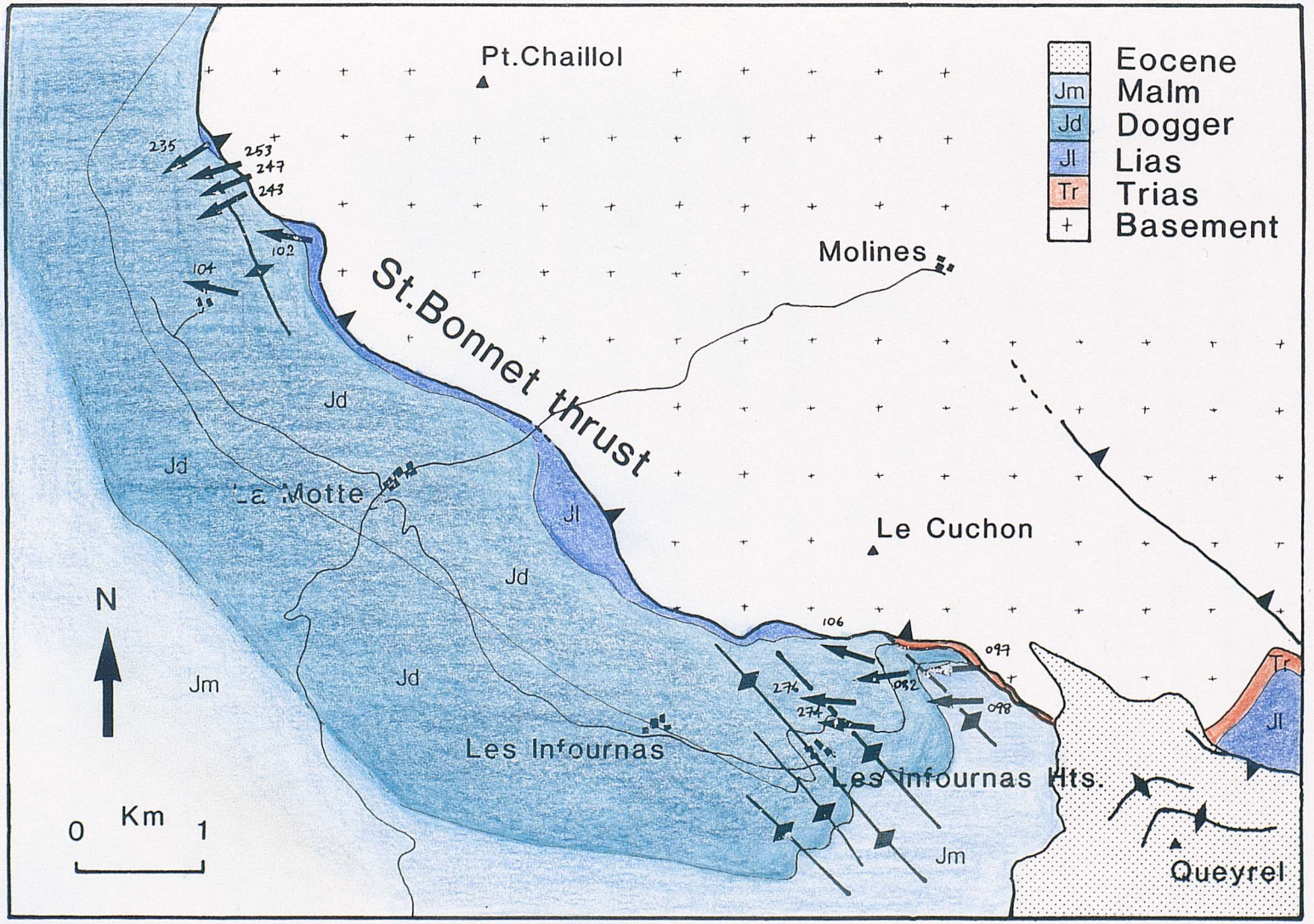


Fig.4.5 MAP OF THE ST.BONNET LOCALITY, SOUTHERN PELVOUX

basement thrust preventing direct juxtaposition of basement with the folded Mesozoic strata. The mapped outcrop describes an E–ENE trending footwall ramp; the thrust cutting up section laterally towards the south as it successively overrides Triassic, Liassic, Dogger and Malm beds.

Further constraint on the transport direction is provided by mineral lineations on the thrust surfaces in basement schists and Triassic spilites. Cleavage within the Lias limestones and shales can also be used as an indicator of the strain ellipsoid. This data yields a mean tectonic transport direction of N260°E with a standard deviation of 20°. Measurements recorded lie in the range N227°E–N286°E. The fault plane itself dips at 35–40°ENE at Mallacua, 2km NW of La Motte. Towards the southeast the fault plane becomes subhorizontal. This is due to the interaction of the lateral thrust ramp and the post–Nummulitic regional tilt of the southern margin of Pelvoux.

The folds in the footwall of the St. Bonnet thrust are tight and overturned. The folds verge towards the W–WSW, but this direction swings towards the southwest as the lateral ramp is approached. The fold axes, as defined by bedding data, plunge at 25–45°SSE–SE. The overturned fold limbs strike NW–NNW and dip at 65–85°NE. The fold train is alone in southern Pelvoux in comprising tight, closely–spaced folds which contrasts with the large scale footwall folds of Muande, Cedera and La Selle. This is because the latter involve basement rocks and the more competent Trias and lower Lias carbonates whilst the former involve the thick shaley succession of upper Lias to Malm age.

The structures of St. Bonnet/Les Infournas Hauts are therefore the result W–WSW–vergent pre–Nummulitic deformation involving basement overthrusting. The fact that the St. Bonnet thrust truncates the folds in its footwall indicates that deformation was progressive and cut–off relationships suggest fault–propagation folding. Early folding and layer–parallel shortening was followed by the propagation of basement thrusts. The presence of Triassic strata in the immediate footwall of the thrust does not indicate minor displacement on the thrust because the Triassic slices are often much sheared and appear to have been carried as thrust horses along the basement thrust plane. The frontal hangingwall ramp is now eroded so exact measurement of the dis–

placement is impossible. Shearing of the basement over the Mesozoic shales produced an intense axial planar cleavage in the footwall strata.

4.1.2. La Muande

This locality lies at the head of the La Muande valley, on the northern flank of a ridge comprising the Queyrel (2440m), Lingustier (2360m) and Tourond (2743m) peaks. Access is gained from the Severaissette valley via la Motte-en-Champsaur and Molines-en-Champsaur. From this point forest paths lead to Le Roy and the Col du Viallet (2240m) and Col de Cote Longue (2679m) (Fig.4.2).

Both pre- and post-Nummulitic age structures have been mapped in the La Muande locality (Fig.4.6.). The former is manifest by WSW-vergent basement overthrusting and folding of the Mesozoic sediments in its footwall. The latter occurred at shallower depths and is typified by shearing in the flysch series.

The major structure at La Muande is the WSW-vergent Vieux Chaillol thrust which carries basement micaschists over the Mesozoic cover sediments. The thrust is truncated by the Nummulitic unconformity west of La Cabane de Vaccivier and is therefore of pre-Nummulitic age. It probably suffered reactivation in post-Nummulitic times since the Nummulitic unconformity is domed at this locality (Fig.4.4). On the slopes of Le Vaccivier, the Vieux Chaillol thrust cuts laterally up-section through the fabric of the basement micaschists. Erosion prior to the deposition of the Eocene flysch series removed the Mesozoic sediments from the hanging-wall of this thrust. They do outcrop, however, to the east in the Tourond valley.

The Vieux Chaillol thrust plane is well-exposed towards the east of La Muande, where it carries basement over overturned lower Lias limestones. As mentioned in section 4.1.1., the Triassic strata, particularly the spilites, tend to form thrust-bounded horses that crop out as pods along the basement thrust plane. Hence, towards the southeastern limit of the Vieux Chaillol thrust outcrop, spilites intervene between the basement and the overturned Liassic sequence.

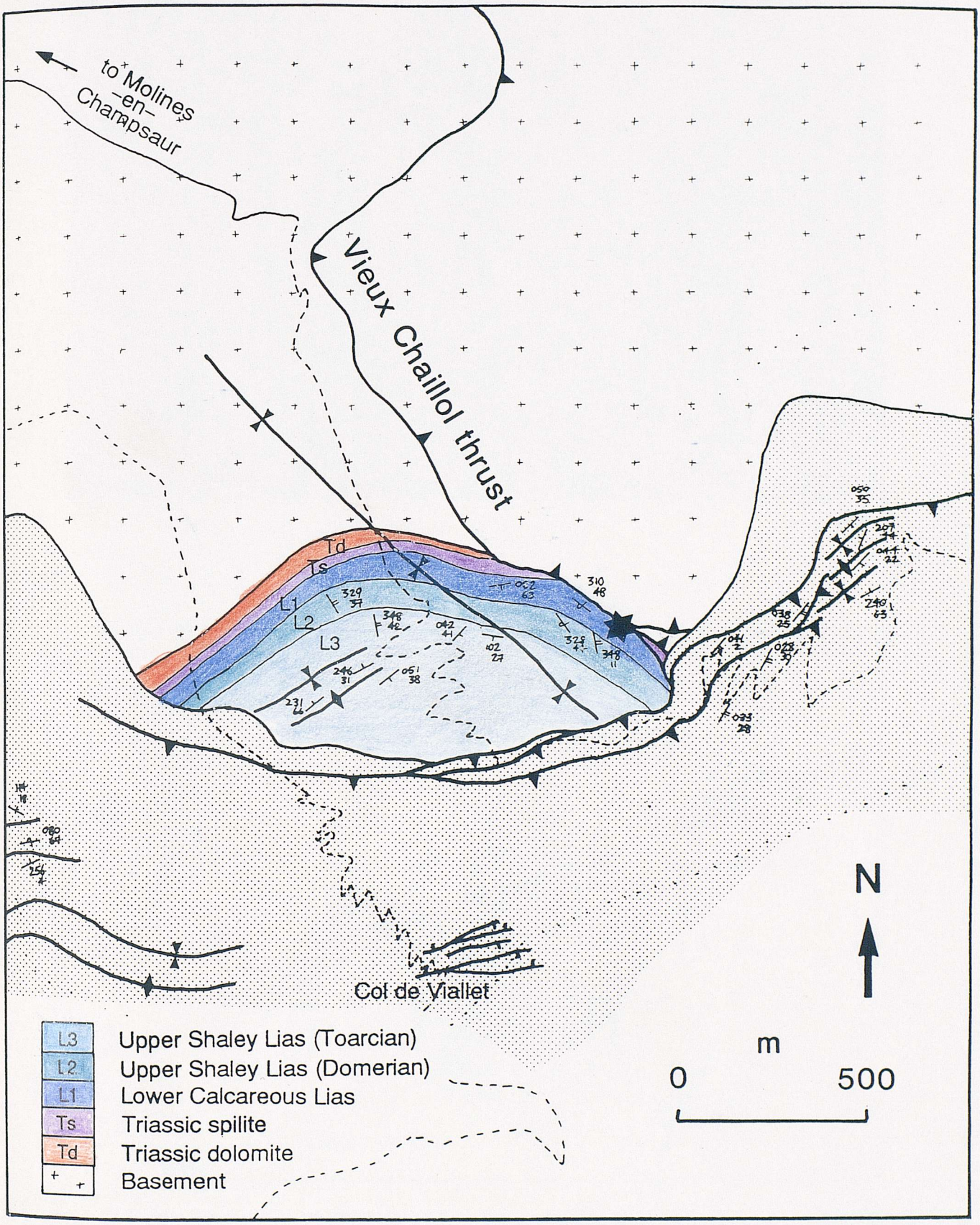
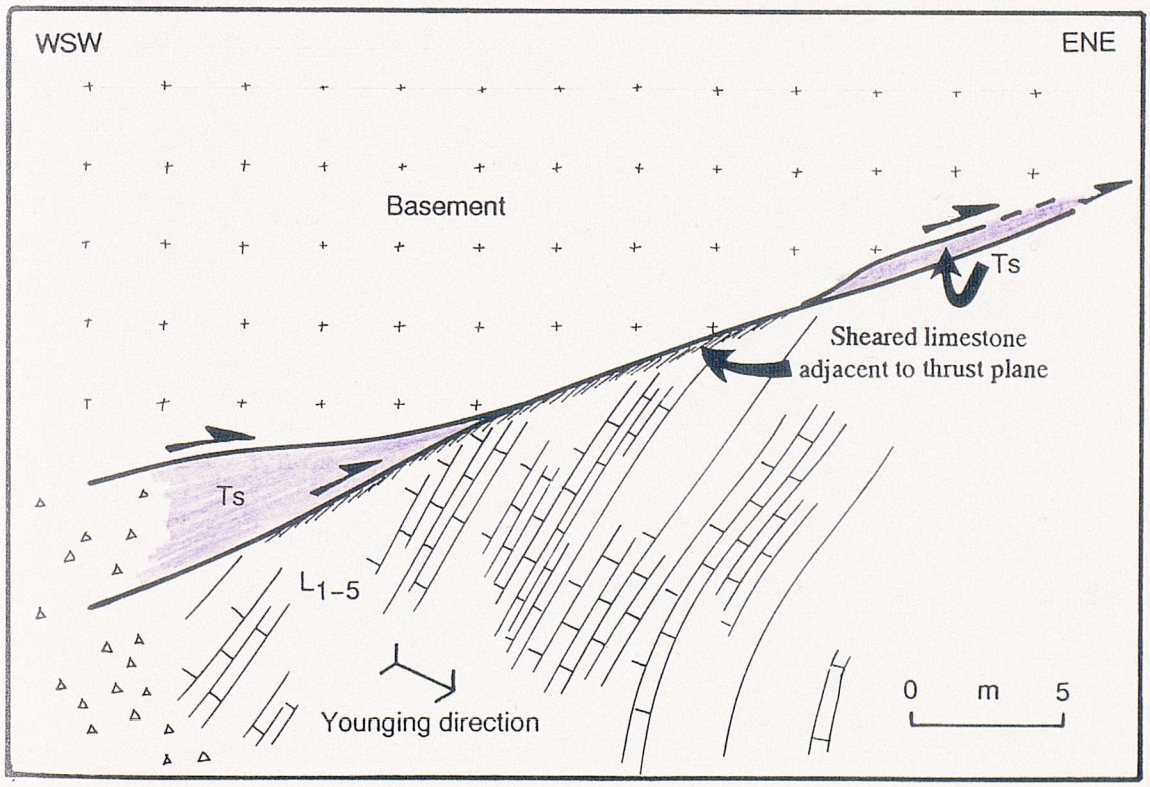


Fig.4.6 MAP OF THE MUANDE LOCALITY, SOUTHERN PELVOUX.





THE OUTCROP OF THE VIEUX CHAILLOL THRUST AT MUANDE,

showing the crystalline basement in the hangingwall juxtaposed against overturned Lower calcareous Lias strata. Note the pods of Triassic spilite outcropping along the thrust plane.
For locaton see Fig.4.6. ✸

In the footwall of the Vieux Chaillol thrust both crystalline basement and Triassic–Liassic sediments are folded into a large, overturned, WSW–vergent syncline (Fig.4.4.). The fold plunges towards the SE–SSE due to the regional tilt imposed during post–Nummulitic uplift. At the thrust contact, the well–bedded lower Lias carbonates are sheared towards parallelism with the thrust plane and a strong, fanning cleavage is developed in the interbedded shales. The over–turned beds strike NW–SE and dip 45–60°NE, before passing through subvertical dips in the fold hinge. In the right–way–up limb bedding dips at 25–40°S–SSE reflecting the regional south–southeastward tilt of the southern margin of the Pelvoux massif; a tilt caused by post–Nummulitic uplift.

Cleavage data from the Liassic strata on the right–way–up limb of the syncline strike consistently NW–SE (N329°–N348°E) and dip at 35–45°NE. The cleavage fans around the fold axis and becomes subhorizontal (dipping 5–11°E) on the overturned limb of the syncline near the Vieux Chaillol thrust plane. **Hence the combined cut–off, fold and cleavage data indicates pre–Nummulitic overthrusting on the Vieux Chaillol thrust towards the WSW.**

At La Muande, the post–Nummulitic phase of deformation is restricted to high structural levels, the main detachment horizon being in the Globigerina marls (Fig.4.7.). However, along strike several small branching thrusts have cut down into the Mesozoic sequence, breaching the Eocene uncomformity. Further east, a number of small thrusts splay from the Globigerina marl into the Gres du Champsaur flysch series. With decreasing displacement the thrusts tip out into folds. The fold axes trend NE–SW and plunge gently southwestward. In the fold limbs, bedding dips at 45–65°NW and 22–35°SE, indicating NW–vergent shear. The cleavage associated with these structures strikes approximately NNE–SSW to NE–SW (N028°E–N041°E) and dips at 25–30°SE.

The post–Nummulitic NW–vergent shear has also produced folds in the Mesozoic strata. On the right–way–up limb of the pre–Nummulitic syncline NE–SW trending folds have been mapped. These are observed to the west of the locality, at Clot Tartaou, where the upper shaley

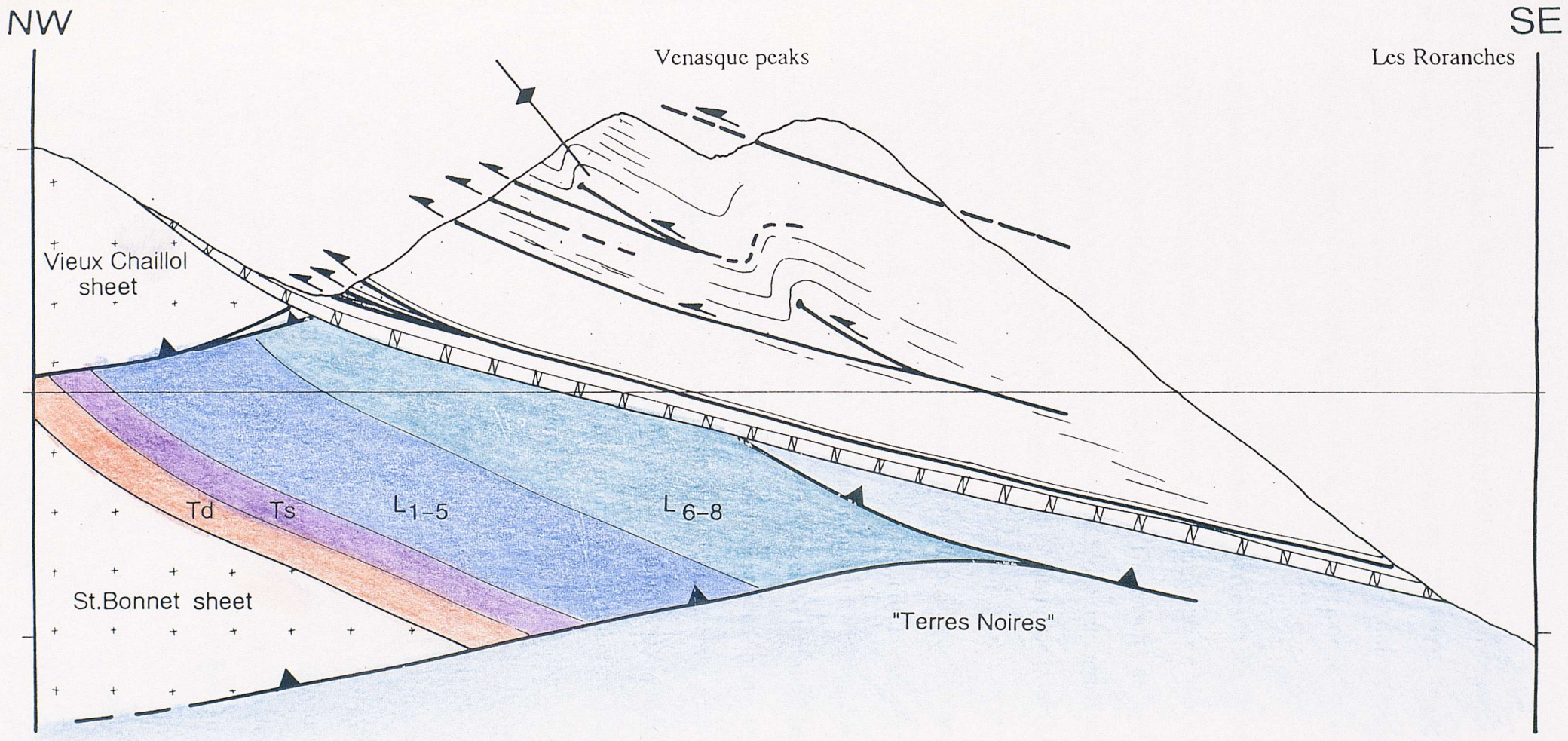


Fig.4.7 SCHEMATIC STRIKE SECTION THROUGH THE MUANDE LOCALITY, showing the lateral ramp terminations of the Pre-Nummulitic Vieux Chaillol and St.Bonnet basement thrusts and the high-level NW-vergent Post-Nummulitic deformation.

Td, Triassic dolomites; Ts, Triassic spilites; L_{1-5}, Lower Calcareous Lias; L_{6-8}, Upper Shaley Lias; N, Nummulitic limestone; dotted area, Gres du Champsaur.

Lias strata are exposed. These incompetent beds are sheared into asymmetric folds with steep limbs dipping 35–65°NW and shallow limbs dipping 10–35°SE. Importantly, the principal cleavage in this area strikes NNW–SSE and is related to pre–Nummulitic compression. A weak NE–SW trending post–Nummulitic foliation is superimposed upon it.

A number of extensional collapse structures are observed in the flysch series on the western flank of Pointe de Lingustier. The faults form a branching array; the individual synthetic and antithetic splays accommodating 2–5m displacement. This results in a total downthrow of 8–10m to the north on the principal fault. The WSW–ENE orientation of these faults suggests that they formed at the same time as the NW–vergent post–Nummulitic overrthrusting.

Further extensional faults have been mapped on the overturned limb of the pre–Nummulitic syncline. These faults offset the lower calcareous Lias and the upper shaley Lias units. They trend approximately E–W and downthrow towards the north by 5–8m. They are related to thrust loading rather than earlier Mesozoic extension. It is suggested that the lateral hangingwall ramp of the Vieux Chaillol thrust caused a greater load to be applied in the north than in the south, generating a set of small offsets in the Liassic strata which trend parallel to the lateral ramp. Alternatively, they may have formed during post–Nummulitic NW–vergent deformation, acting as Riedel–type shears beneath the main post–Nummulitic Globigerina marl–hosted detachment. Such structures have been previously described in the external French Alps (Gidon, 1988 & 1989).

4.1.3. Tourond and Cedera

Southern Pelvoux is drained by a major river, Le Drac, which divides into two tributaries near its source, Le Drac noir and Le Drac blanc or Drac de Champoleon. In the Champoleon valley, at the village of Les Gondouins, two outcrops of Mesozoic strata are sandwiched between the Hercynian basement massif and the Eocene flysch series. One lies to the west, in the Tourond valley and is reached by footpath from Les Fermons. The other is visible from Les Gondouins

and lies on the flank of the Aiguille de Cedera. The summit of the Aig. de Cedera is reached via the Meollion valley in the south (Fig.4.8).

Two major basement thrusts are exposed in this locality, the Tourond and Cedera thrusts (Fig.4.8). These structures verge W–WSW and are unconformably draped by the Eocene flysch trilogy. Hence they are of pre–Nummulitic age. In Tourond a complex linked thrust pattern has been mapped in the thin Mesozoic cover sequence that lies in the footwall of the Tourond thrust. Later NW–vergent post–Nummulitic deformation has imbricated and folded the Eocene flysch series. In addition, the underlying Triassic strata have been affected by the superposition of post–Nummulitic deformation upon the pre–Nummulitic structures. Hence, early WSW–vergent thrusts have been reactivated in a NW–vergent sense and new breaching faults have formed. Locally these thrusts breach the Eocene unconformity.

South of the Tete de Mal Cros (3086m) summit the granitic basement of the Tourond thrust sheet overrides the schistose basement of the Vieux Chaillol sheet (Figs.4.10a&b). Due to the present day erosion level, much of the Tourond locality represents a structural window through which the Triassic cover in the footwall of the Tourond thrust can be seen. These sediments are imbricated and folded on a small scale; the grey–white marls of the lower Triassic acting as a detachment horizon.

In the centre of the Tourond locality the Vieux Chaillol basement is exposed in two windows lying along the axis of a NNW–SSE trending anticline (Fig.4.8). Imbricates of Triassic strata are folded about this axis indicating that this domal uplift was a later development. Two imbricate stacks representing unroofed duplexes have been mapped. One outcrops to the west of the Chalets du Tourond (Fig.4.10b) and the other, further north, to the west of the Puy des Pourroys (2784m) (Fig.4.10d).

The southern stack formed by footwall collapse during the propagation of the Tourond basement thrust (Fig.4.10b&c). In the immediate footwall of this thrust, at the chalets, lies a

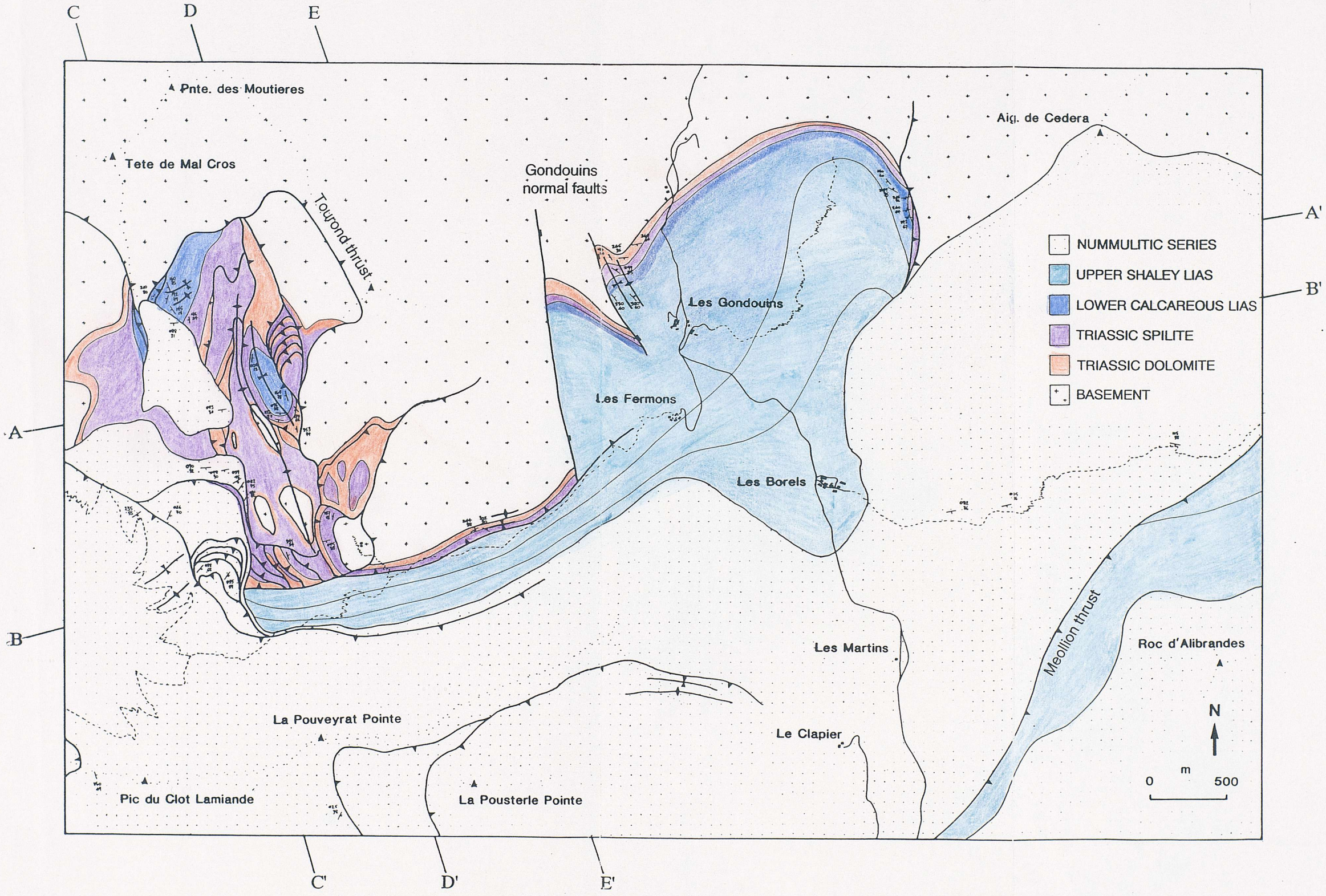


Fig.4.8 MAP OF THE TOUROND AND CEDERA AREAS, SOUTHERN PELVOUX.

Sectons A–A', B–B', C–C', D–D' & E–E' are shown in Figures 4.10a,b,c,d & e respectively.

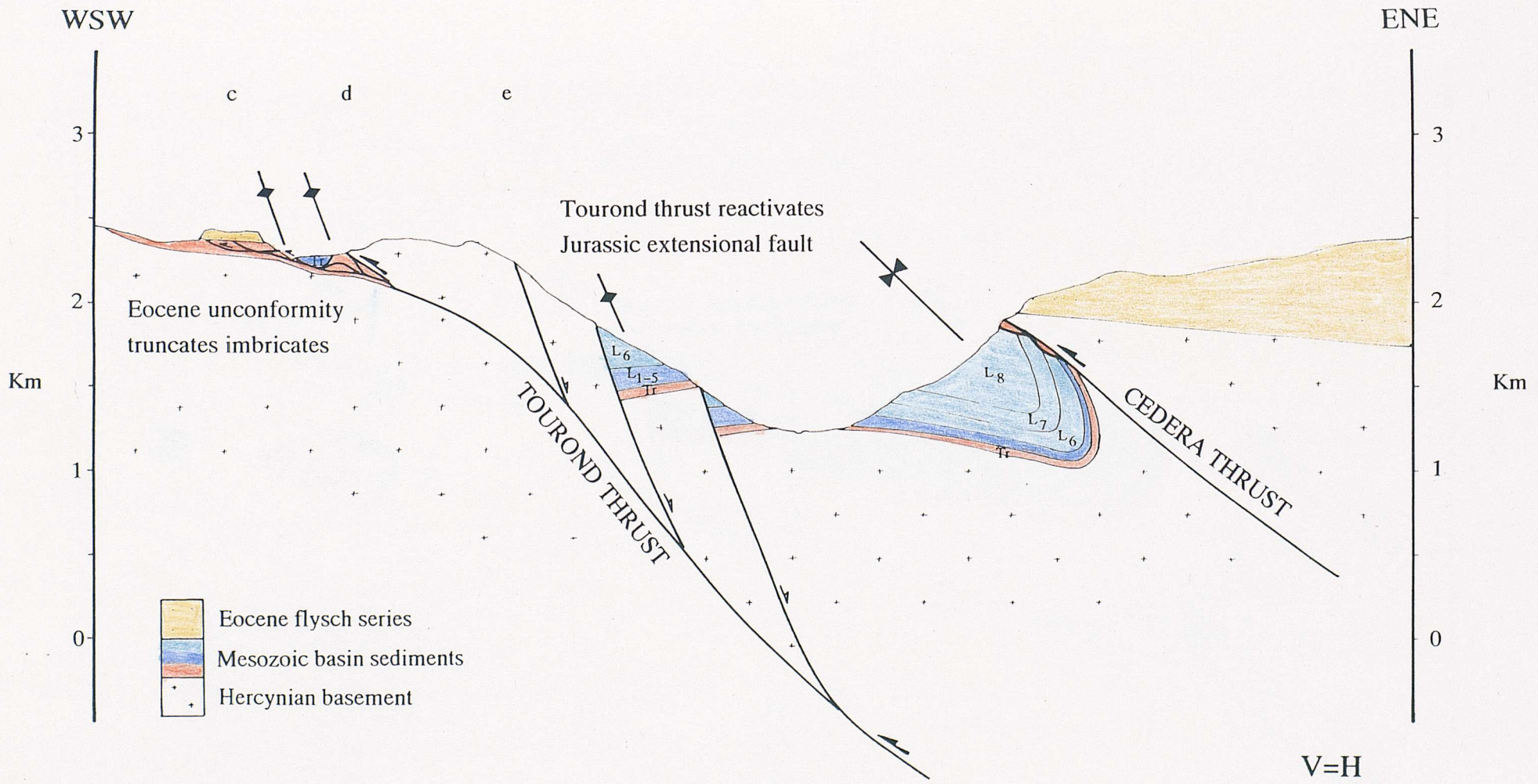


Fig.4.10a. SECTION ACROSS TOUROND– CEDERA STRUCTURE,

showing two styles of thrusting during pre–Nummulitic deformation;
1) the reactivation at depth of the Gondouins normal fault (Tourond T.) and
2) "pli–faille" breach thrusting (Cedera T.)

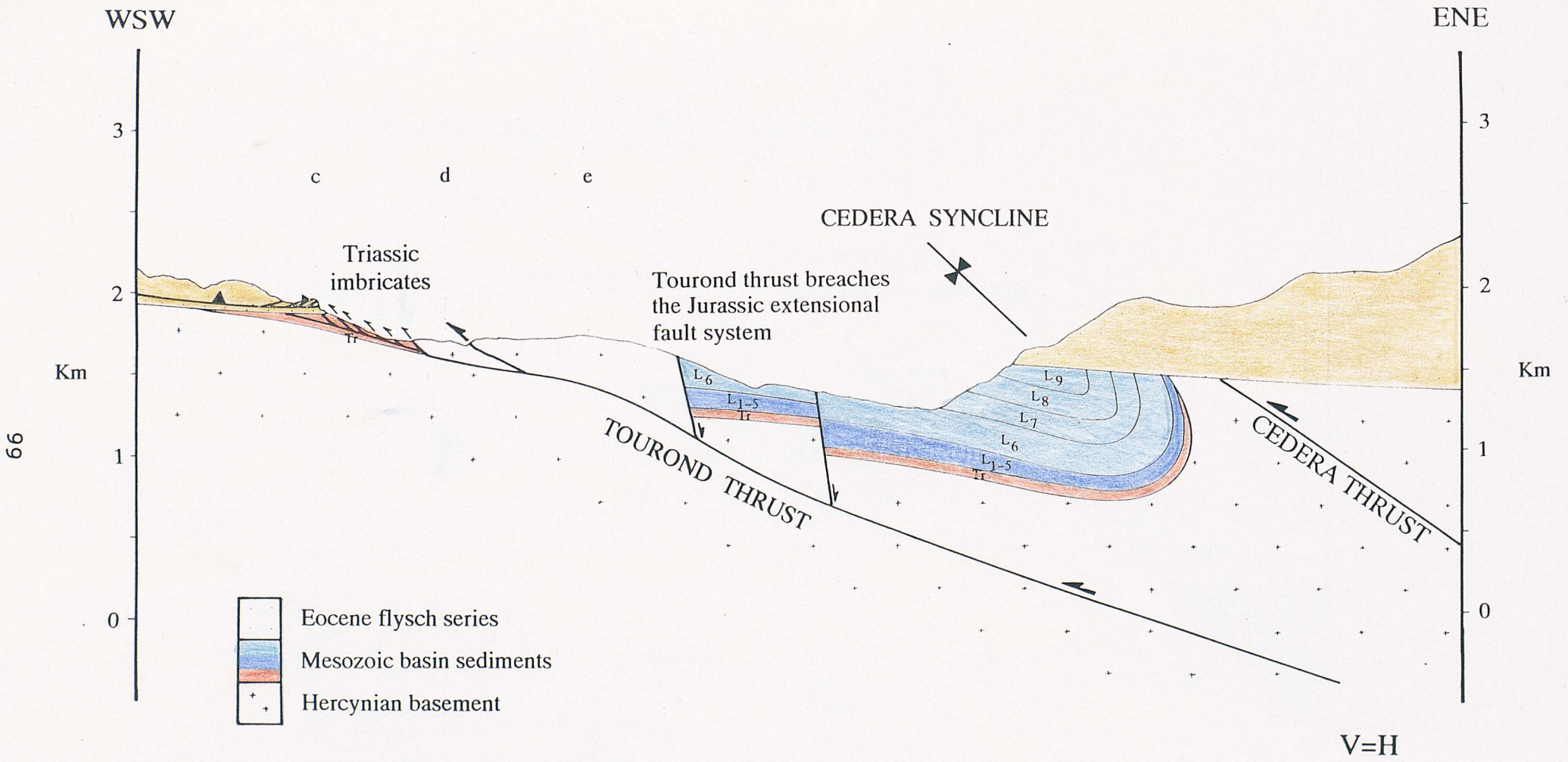


Fig.4.10b. SECTION ACROSS TOUROND–CEDERA STRUCTURE,

showing the Tourond thrust adopting a low angle breaching geometry
as it ramps laterally to the south

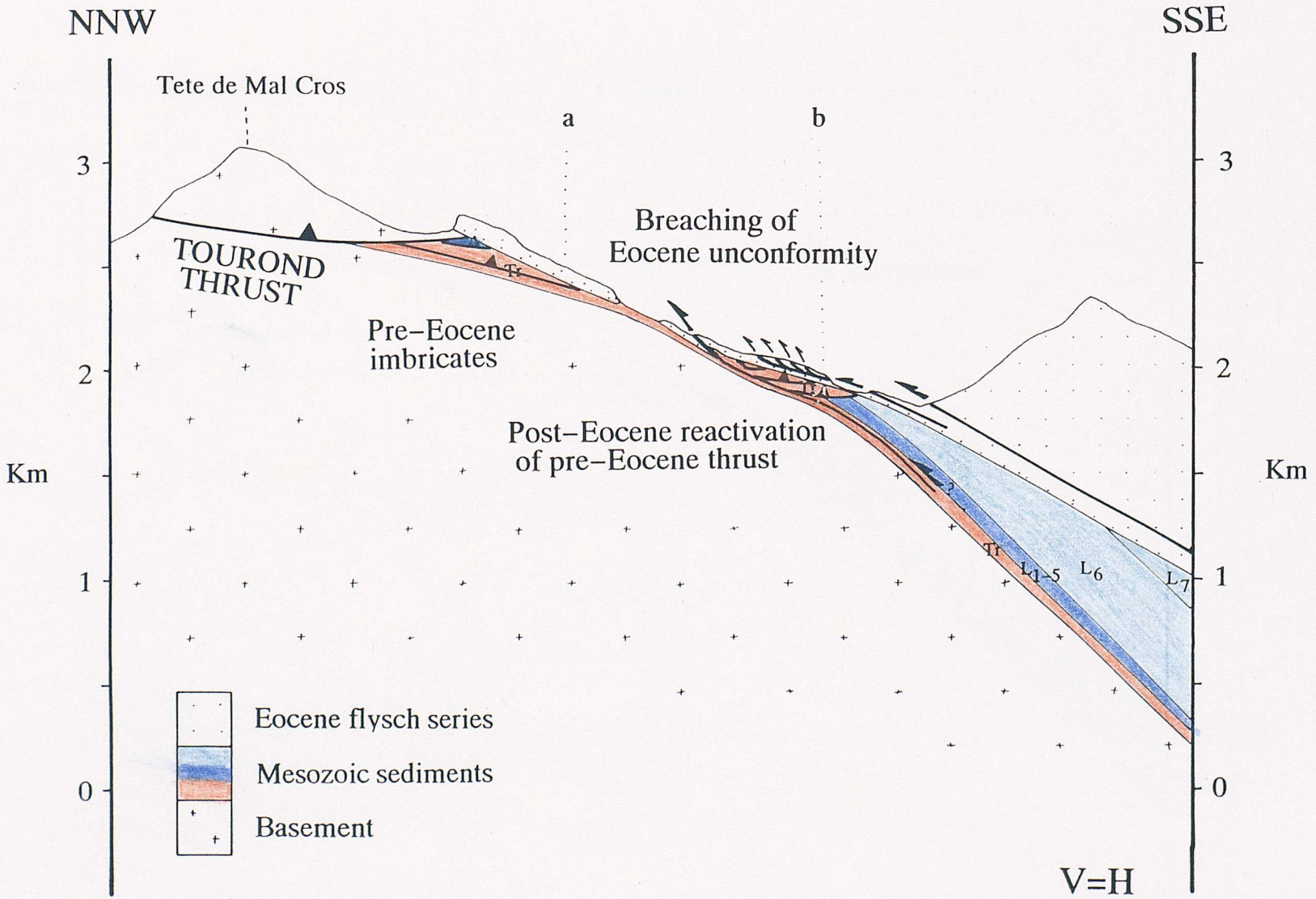


Fig.4.10c. SECTION ACROSS TOUROND STRUCTURE,
drawn sub–parallel to the post–Nummulitic transport direction

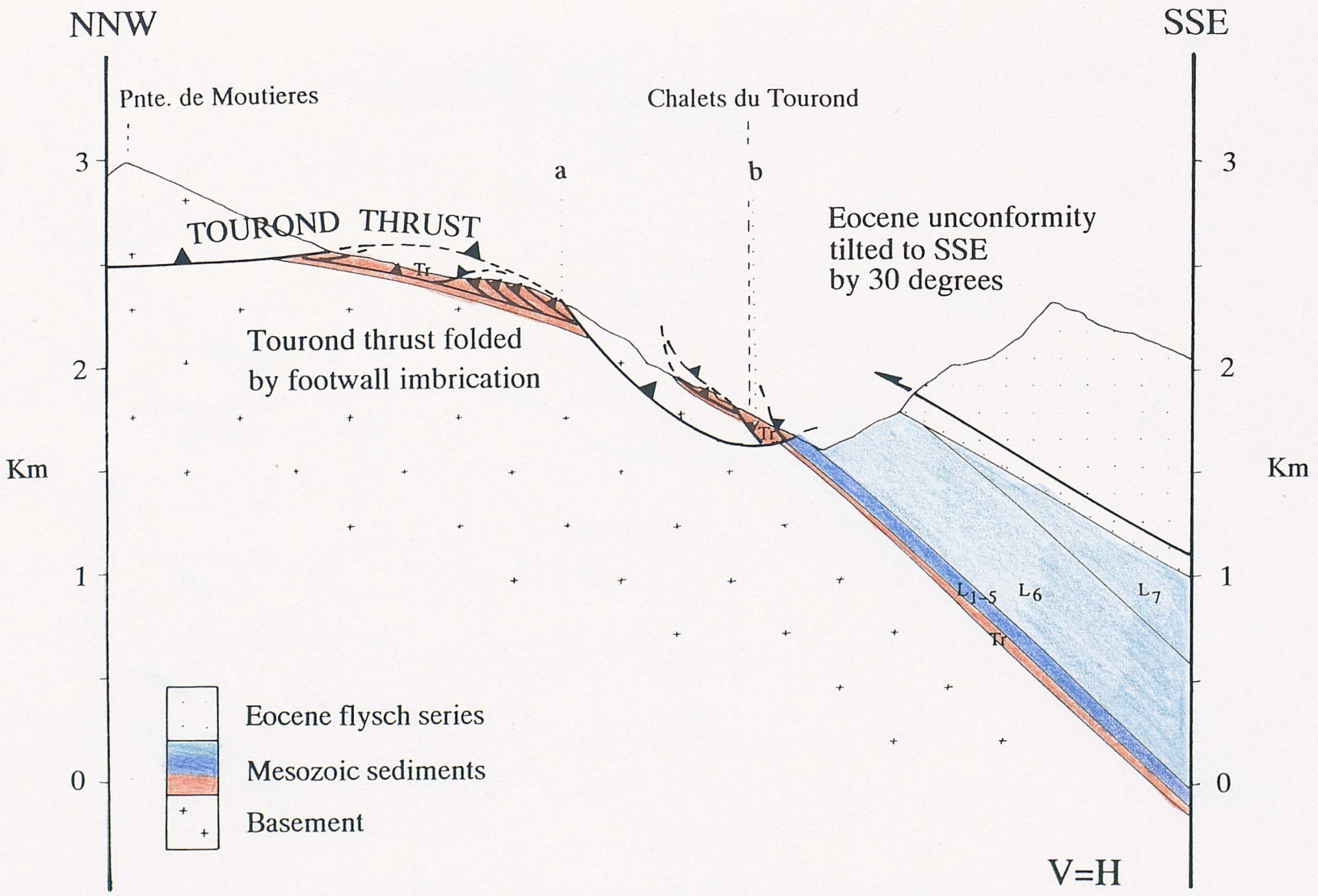


Fig.4.10d. SECTION ACROSS TOUROND STRUCTURE

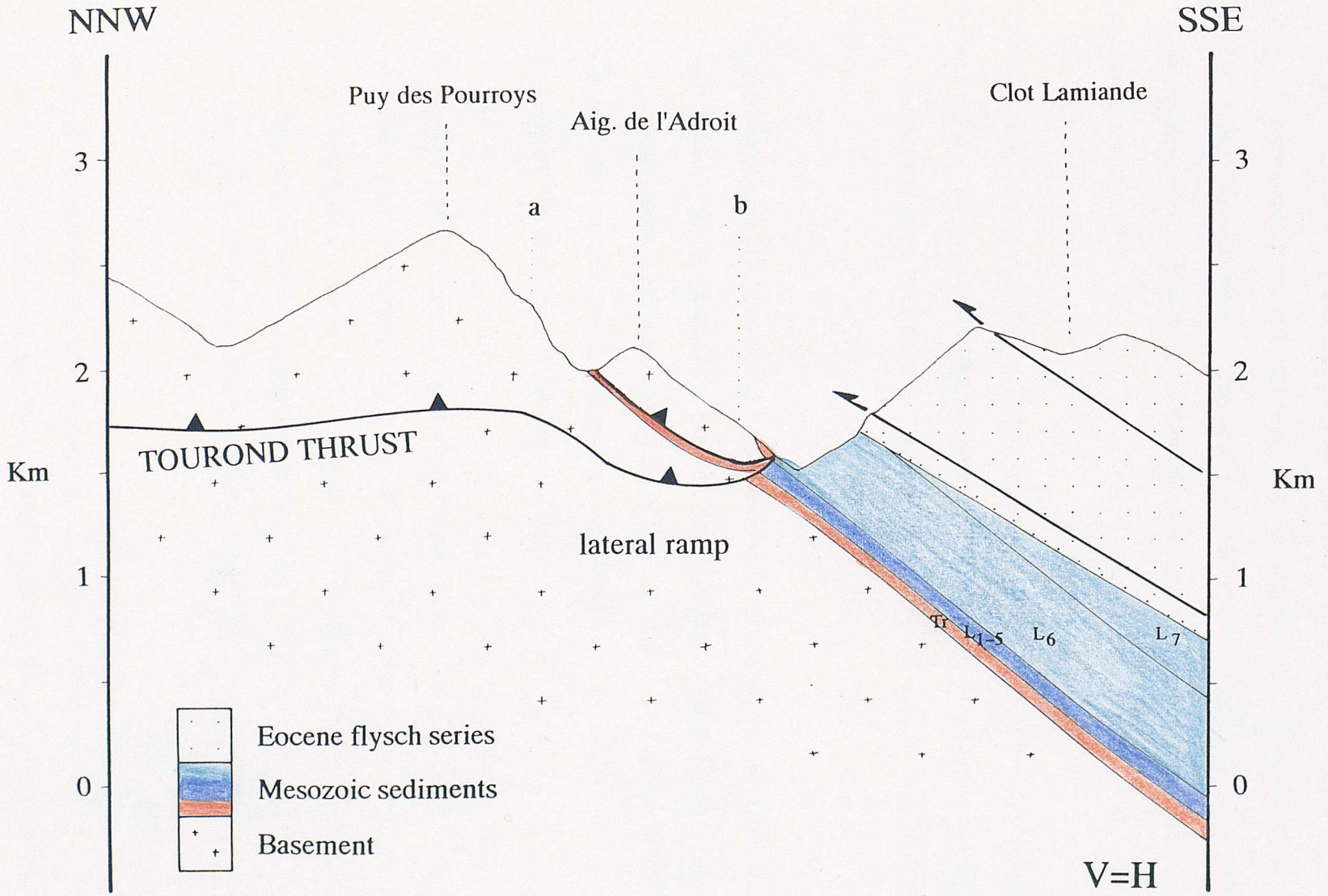


Fig.4.10e. SECTION ACROSS TOUROND STRUCTURE,
showing lateral ramping of pre–Nummulitic basement thrust sheets.

small basement horse with Triassic cover sediments preserved in its frontal hangingwall ramp. This thrust sheet overrides 5 or 6 small Triassic imbricates. Mineral lineation data recorded from the Triassic spilites indicate a predominant WSW–W vergent (N245–275°E) shear which is supported by measurements from the basement horse in the footwall to the Tourond thrust (N246–271°E). During post–Nummulitic compression the floor thrust of the stack was reactivated and Triassic spilites were carried over the Nummulitic limestone member of the Eocene trilogy, breaching the unconformity.

In the northern imbricate stack (Fig.4.10d), which outcrops as a window, the Triassic dolomites and spilites are repeated 5 times. Lateral thrust ramps that formed during pre–Nummulitic deformation have been reactivated as frontal structures during post–Nummulitic compression.

Lower Lias sediments are preserved in a syncline core at Tourond. This syncline plunges SSE due to the regional tilting of southern Pelvoux and is carried westwards on a late pre–Nummulitic breaching thrust. The breaching nature of this structure can be seen on the map. Otherwise, Liassic strata are noticeably absent from the locality – a feature attributed to the development of a thrust flat at the base of the lower calcareous Lias unit.

The lower calcareous Lias strata, comprising a thinly–bedded alternating limestone and shale sequence, also outcrops immediately below the Eocene unconformity at the Crete de l'Arche. The strata are folded about W–WNW vergent, SSW–plunging axes. Bedding strikes at N190°–210°E in the steeply WNW–dipping (68–90°) limbs and strikes at N145°–150°E in the shallowly SE–dipping (18–25°) limbs. These folds are interpreted as fault–bend folds related to lateral ramps.

Two features characterise the pre–Nummulitic structure of Tourond :–

1) The first is the exclusion of Liassic sediments from the thrust imbricates. The vast majority of the thrust slices comprise basement and Triassic strata alone, the Triassic dolomites

and spilites being frequently imbricated together without the involvement of the overlying Liassic sediments.

2) The second is the large scale lateral ramp that dictates the southern margin of the Pelvoux massif. This ramp is observed at St. Bonnet, La Muande and Tourond (Fig.4.7). The basement thrusts cut up-section to Jurassic horizons towards the south.

These two observations constrain any model proposed for the region. The proposed thrust propagation sequence is illustrated in Figure 4.11.

1) The sequence begins with the ramping of the Tourond thrust through the basement and Triassic cover before becoming bedding parallel at the Triassic-Liassic boundary. Hence the basement was displaced approximately 750-1000m across this flat and the majority of the Liassic strata was carried away at the toe of the Tourond Thrust, leaving only Triassic strata in its footwall. The continuation of this thrust at depth is unknown, but it may steepen and reactivate the Gondouins fault (Fig.4.10a) or follow a shallowly-dipping breaching geometry (Fig.4.11.). In either case the Tourond thrust must ramp laterally towards the south as it rises into the Oxfordian "terres noires" shales (Figs.4.10a&b).

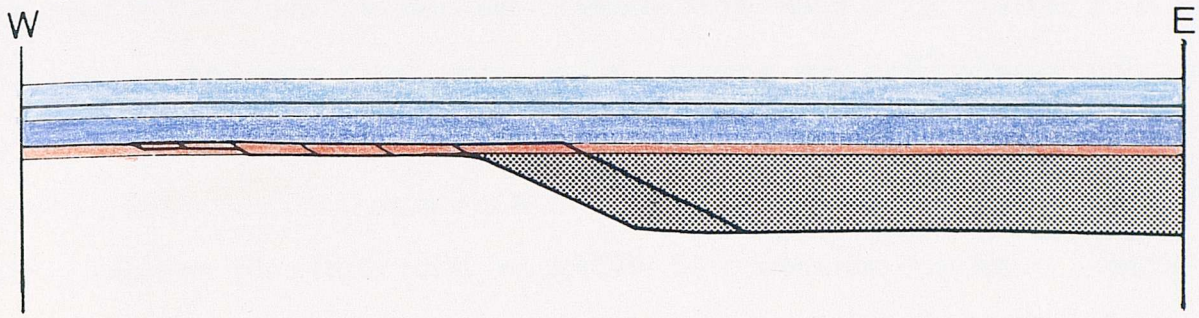
2) Further displacement was accommodated by imbrication of the Triassic strata in the footwall of the Tourond thrust.

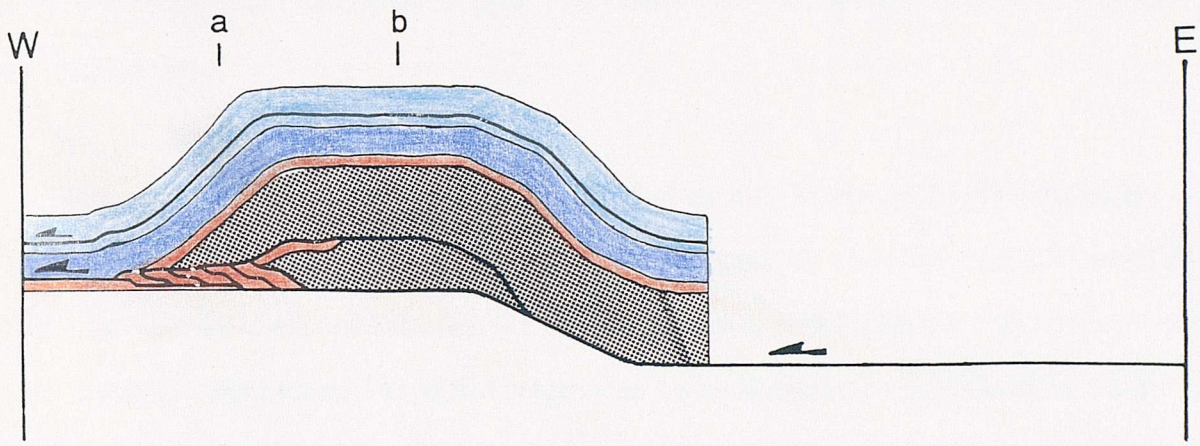
3) Then displacement was transferred to the Vieux Chaillol thrust.

4) The "terres noires" detachment is inferred to have extended northwards across the Pelvoux massif, partitioning deformation in the basement-Liassic strata and the Dogger-Cretaceous strata.

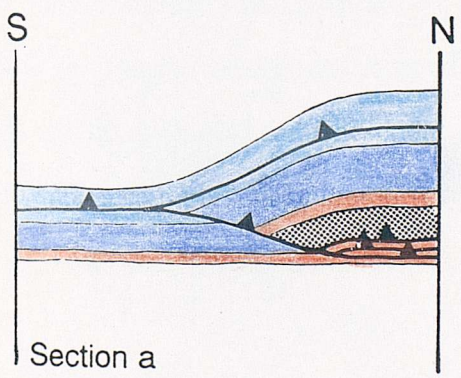
Post-Nummulitic deformation of the Nummulitic trilogy is well-exposed to the south of Les Fermons, at the Croix de l'Estang (2209m). Shears splay off a detachment in the Globigerina marl unit of the Nummulitic trilogy and die as large tip folds in the Gres du Champsaur flysch

1) Development of footwall flats at base Triassic and base Liassic horizons.





2) Development of a lateral culmination above lateral ramp stack.



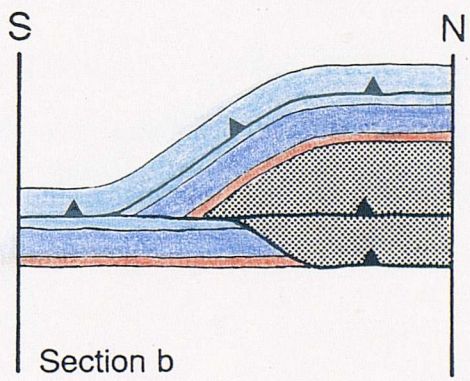


Fig.4.11. CARTOONS SHOWING GEOMETRY OF THE TOUROND THRUST AND ITS FOOTWALL IMBRICATES.

series. Kinematic indicators, particularly mineral lineations (Fig.4.8), indicate that the deformation was WNW–NW (N134°–141°E) vergent. This post–Nummulitic compression direction is confirmed in the Nummulitic limestone exposures to the north of the Cascade de la Pisse (Fig.4.10c). At this locality detachments at the basal conglomerate–Nummulitic limestone interface and within the Globigerina marl unit produce a duplex comprising horses of Nummulitic limestone. Shear lineations (azimuths of N124°–148°E) and cleavage foliations (striking N026°–038°E) within the Globigerina marls indicate NW–SE compression. In addition numerous en echelon tension gashes occur in the Nummulitic limestone and these trend N135°–149°E, indicating a **NW–SE post–Nummulitic principal compressive stress** consistent with the cleavage and lineation data.

The post–Nummulitic deformation occurred at shallow structural levels. The phase did, however, affect the Mesozoic sequence locally. At the Cascade de la Pisse duplex (Figs.4.8 & 4.10c) just described, a thrust breaches the Nummulitic unconformity and carries Triassic spilites over Nummulitic limestone. This thrust reactivates a pre–Nummulitic thrust plane. Elsewhere in the Tourond locality there is evidence of the post–Nummulitic reactivation of pre–Nummulitic structures, particularly the lower Triassic grey marly–shale detachment. Data from this slip plane record two distinct sets of lineations, one WSW–vergent and the other NW–vergent. East of the Crete de l'Arche lineations on small–scale, subvertical, NNW–SSE trending shears indicate NW–NNW vergent shear. These late shears do not map out to form a linked thrust system; rather they appear to represent scattered fractures in the footwall to the main post–Nummulitic detachment (within the Globigerina marl?) which has since been eroded.

The Tourond valley is bounded to the south by several peaks, Pointe sud de la Venasque (2620m), Pic du Clot Lamiande (2461m), La Pouveyrat Pointe (2467m) and La Pousterle Pointe (2401m). Further to the south lies La Palastre (2279m). These hills are composed of Gres du Champsaur flysch sediments for the most part, but their summits are composed of klippen of middle–upper Jurassic marly limestones and Eocene Nummulitic limestone.

The klippen are related to the post–Nummulitic age Sirac basement thrust which carries granitic gneisses of the Hercynian basement suite over the Eocene flysch at Rougnoux (section 4.2.1). Moving westward to the Meollion valley and then Tourond/Soleil Boeuf, successively younger hangingwall strata are juxtaposed against the flysch series in the footwall – describing a frontal hangingwall ramp and constraining the transport direction.

At La Palastre, the Nummulitic limestone has been carried across the Gres du Champsaur flysch developing a strong cleavage which strikes N310°–346°E and dips moderately at 53–65°E. Lineations are oriented N233°–256°E. This correlates with a principal stress direction of N076°E indicated by tension gashes at La Pousterle Pointe and transport direction indicators from the footwall to the Sirac thrust at Col du Cheval de Bois. Hence, **the combined kinematic evidence indicates that the post–Nummulitic Sirac thrust system was WSW–directed**.

This contrasts with the NW–vergent post–Nummulitic deformation observed at La Muande and in the Tourond valley where the base of the Eocene sequence is observed. This indicates that the post–Nummulitic transport direction varied from NW to WSW with decreasing structural level. This may reflect :–

1) the constraint imposed by the basement topography that was generated during pre–Nummulitic deformation and may have inhibited post–Nummulitic shearing. This created anomalous transport directions in close proximity to basement highs whilst higher in the flysch pile the true transport direction is recorded. **On a more regional scale, kinematic indicators from the Arvinche, Dourmillouse and Fournel show that the post–Nummulitic transport direction was W–WSW directed**.

2) Temporal variation in post–Nummulitic transport direction. Lineations at Pic du Clot Lamiande and La Pouveyrat indicate NW–vergent post–Nummulitic shear. These localities lie in close proximity to the WSW–directed Sirac–related klippen indicating that the boundary between the two shear sets is abrupt, not transitional. The late post–Nummulitic Sirac structures

have breached through the earlier post–Nummulitic structures.

In the east, at the Cedera locality, the basement and Mesozoic strata are folded into a large SW–vergent syncline (Fig.4.10a&b) that has been breached by a late stage thrust. This structure is also truncated by the Eocene unconformity. Its structural style closely resembles that of the La Muande fold–thrust pair.

From the Champoleon valley the basement–cover unconformity is seen to rise steeply from the valley floor to Le Peyron (1981m). The Triassic strata outcrop discontinuously and have been sheared during the final stages of folding. The Liassic and early Dogger sediments form a more continuous outcrop that describes a large fold (Debelmas, Pecher & Barfety, 1989). However, in the east, the lower calcareous Lias unit has also suffered from shearing.

The late shearing breaches the overturned limb of the fold. The Cedera thrust carries basement over the folded Jurassic strata. A number of anastomosing shears in its footwall define pods of Triassic spilite and lower Lias carbonates that have been carried along the basement thrust plane. Pre–Nummulitic erosion has made estimation of the original size of this basement thrust sheet impossible to constrain.

The Cedera fold possesses an axial cleavage striking NW–SE (N299°–310°E) which dips 35°NE. The overturned Liassic beds in the immediate footwall of the breaching thrust strike E–W (N260°–272°E) and dip gently at 25–35°N. The bedding becomes subvertical at the axis and then swings from strikes of SE–NW to E–W on the right–way–up limb. The fold axis therefore plunges 35°SE.

NW of Les Gondouins a set of Mesozoic extensional faults are preserved. These NNW–trending faults downthrow basement and Triassic–Liassic age cover sediments to the east. Two faults are mapped (Fig.4.8), the downthrow on the major fault is approximately 750–1000m and that of the smaller synthetic fault is approximately 250–450m (Fg.4.10a&b). Hence the total

basement offset may have exceeded 1km initially. During later compressional deformation the footwall basement blocks have acted as buttresses to WSW–SW vergent shear in the cover sediments, producing tight folds and an intense cleavage oriented NW–SE. The cleavage fans about the fold axes; striking between N299°E–N320°E and dipping 35–60°NE. They resulted from SW–NE directed compression and could have formed during either the pre– or post–Nummulitic deformation phase. The normal fault planes have not been reactivated, at surface, due to their steep dips. Instead basement thrusts, that may reactivate the deeper portions of the normal faults, have breached the uplifted shoulder of the basement fault block to form "iles flotantes" (Fig.4.10a).

4.2. FIELD DATA FROM SOUTHEASTERN PELVOUX.

The most important mappable structure of southern Pelvoux is the Sirac thrust, which is part of a major post–Nummulitic breaching thrust system. The structures in the footwall of this system were considered in section 4.1., here the Sirac system and the structures in its hangingwall are described. **These structures outcrop at four localities (Fig.4.13) :–**

1) **Meollion/Rougnoux** (section 4.2.1),

2) **Selle** (section 4.2.2),

3) **Fournel** (section 4.2.3) and

4) **Dourmillouse** (section 4.2.4).

The Sirac structure can be traced from the Rougnoux valley westward into the Meollion valley. In this direction progressively younger hangingwall rocks are juxtaposed against the Eo–Oligocene flysch (see section 4.2.1). The mineral lineation and bedding cut–off data recorded in Rougnoux and Meollion indicate a tectonic transport direction towards the WSW (N250°E).

At Rougnoux, pre–Nummulitic deformation is also preserved. Imbricated Triassic strata are truncated by the unconformable Eocene basal conglomerate. Basement faulting is also observed; the basement being downthrown to the SE by a steep, planar fault. This extensional struc–

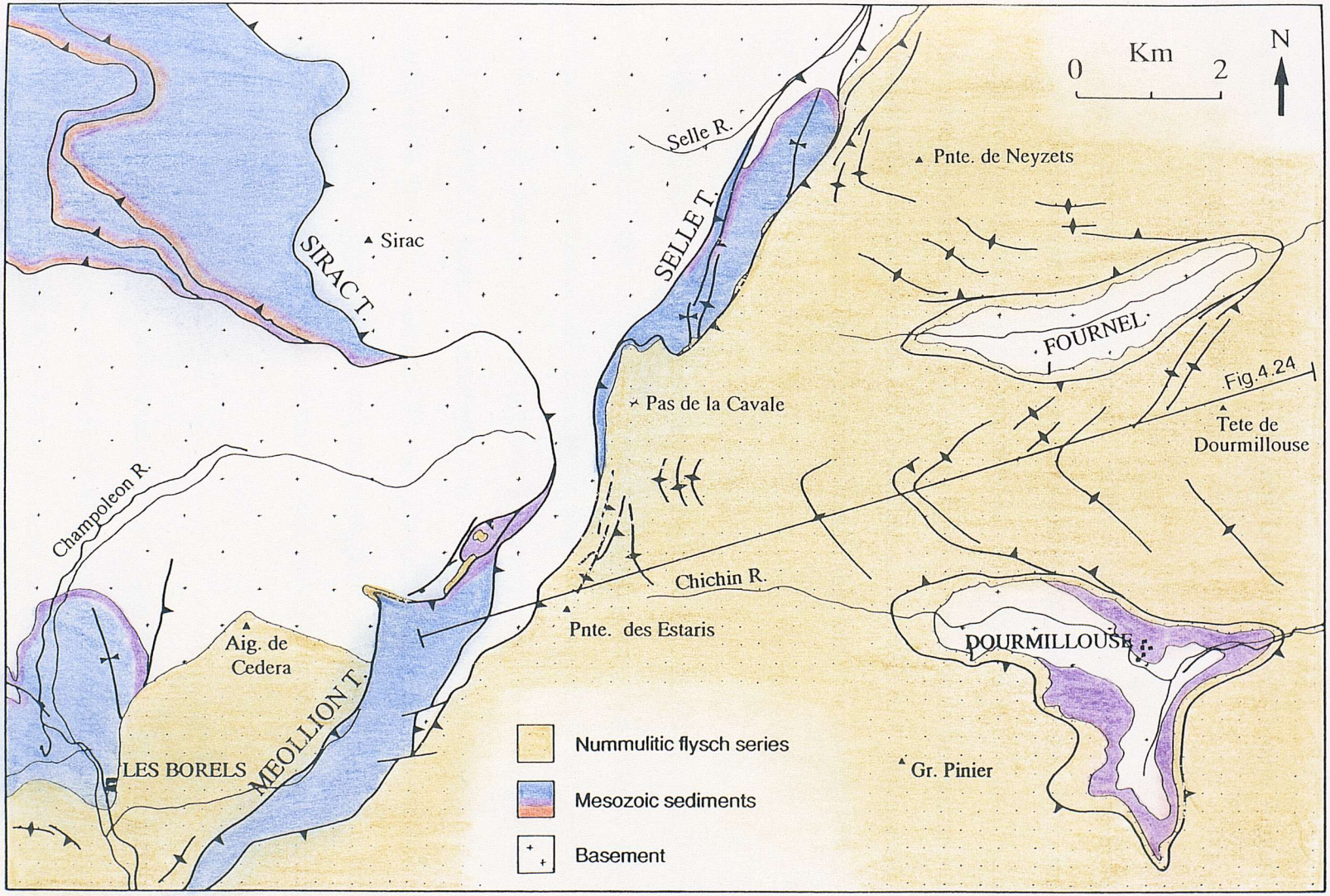


Fig.4.13. MAP OF SOUTHEASTERN PELVOUX,
showing the locations of the Meollion/Rougnoux, Selle, Fournel and Dourmillouse outcrops.

ture is of post–Triassic age since, at outcrop, it displaces both Hercynian basement and Triassic dolomites and spilites. Erosion prevents observation of the relationship between this fault and the Jurassic strata; however comparison with the Les Gondouins faults would suggest an upper Liassic–lower Dogger age of fault movement.

The cover sediments in the hangingwall of the Sirac thrust outcrop in the Selle valley where they have been deformed by both pre– and post–Nummulitic phases of deformation (see section 4.2.2). Because the Eocene unconformity has been sheared at this locality, distinction between the phases is more ambiguous. The style of pre–Nummulitic deformation resembles that of the structures at St. Bonnet, Muande and Cedera being characterised by large scale folding of both basement and cover strata. The overturned limb of this fold is unconformably overlain by the basal Eocene units.

Post–Nummulitic deformation dominates the Selle locality. The flysch trilogy is strongly affected; the Globigerina marl acting as a major decollement from which thrusts splay into the Gres du Champsaur flysch unit and terminate as large scale tipfolds. Unusually, the Eocene unconformity and the Nummulitic limestone unit are also sheared and folded between Pointe de Neyzets and Col de l'Aup Martin. Further west, the Agnel thrust and the Selle thrust in its foot–wall, ramp upwards from the basement–Triassic interface to the Eocene flysch series. These thrusts flatten into the Globigerina marl unit and hence were active in post–Eocene times. Their geometry suggests that they may have originally been pre–Eocene out–of–syncline thrusts that were subsequently reactivated.

The thick Nummulitic flysch series outcrops over a large area to the southeast of Pelvoux. Two windows occur in this area, at Fournel and Dourmillouse, which expose basement and Triassic sequences (see sections 4.2.3. & 4.2.4). At these localities the style of post–Nummulitic deformation in the Eocene strata is well–exposed. The thick succession of alternating shales and sands deform by layer–parallel shearing; shear being concentrated in the shales. Where the shears ramp up section large fault–tip folds are generated and at Fournel spectacular fold trains

are observed. The progressive nature of the post–Nummulitic shearing means that early folds and fault–planes are refolded or breached by later structures. Shear lineations and fold orientations in the flysch indicate WSW–vergent shear. This is consistent with the transport direction deduced for the Sirac thrust system but contrasts with the NW–vergent post–Nummulitic shear that characterises the Tourond and Muande localities.

The structural geology of the four localities (Meollion/Rougnoux, Selle, Fournel and Dourmillouse) will now be described in some detail.

4.2.1. Meollion/Rougnoux

Both of these valleys drain into the Champoleon valley of southern Pelvoux. The Rougnoux valley joins the Champoleon at its eastern limit and is reached via the Cabane Pastorale de Champoleon C.A.F., from where one ascends towards the Col du Cheval de Bois (2473m). The Meollion valley branches eastwards from the Champoleon valley at the village of Les Borels and rises to the Col du Méollion (2451m). These two cols are separated by the small Prelles hanging valley (Fig.4.14).

The Méollion and Rougnoux valleys trend NE–SW. On their southeastern flank lies the thick Gres du Champsaur flysch series which caps a ridge comprising the Pnte. des Pisses (2837m), the Sommet de Prelles (2953m), the Pnte. des Estaris (3086m) and the Pntes. de Rougnoux (3179m). To the northwest lie the basement outcrops of the Aiguille de Cedera (2883m) and Rouite (2663m), which are capped unconformably by Eocene sediments. Sandwiched between these ridges is the Sirac thrust system, comprising the Sirac basement thrust and imbricates of Triassic–lower Jurassic cover in its footwall, principally the Méollion thrust. At La Palastre and Soleil Boeuf, 10km to the west of Col de Méollion, klippen of the Sirac thrust system carry middle and upper Jurassic marls, limestones and shales over the Eocene flysch. Hence the Sirac thrust system is a major breaching system that rises through the stratigraphic section towards the west.

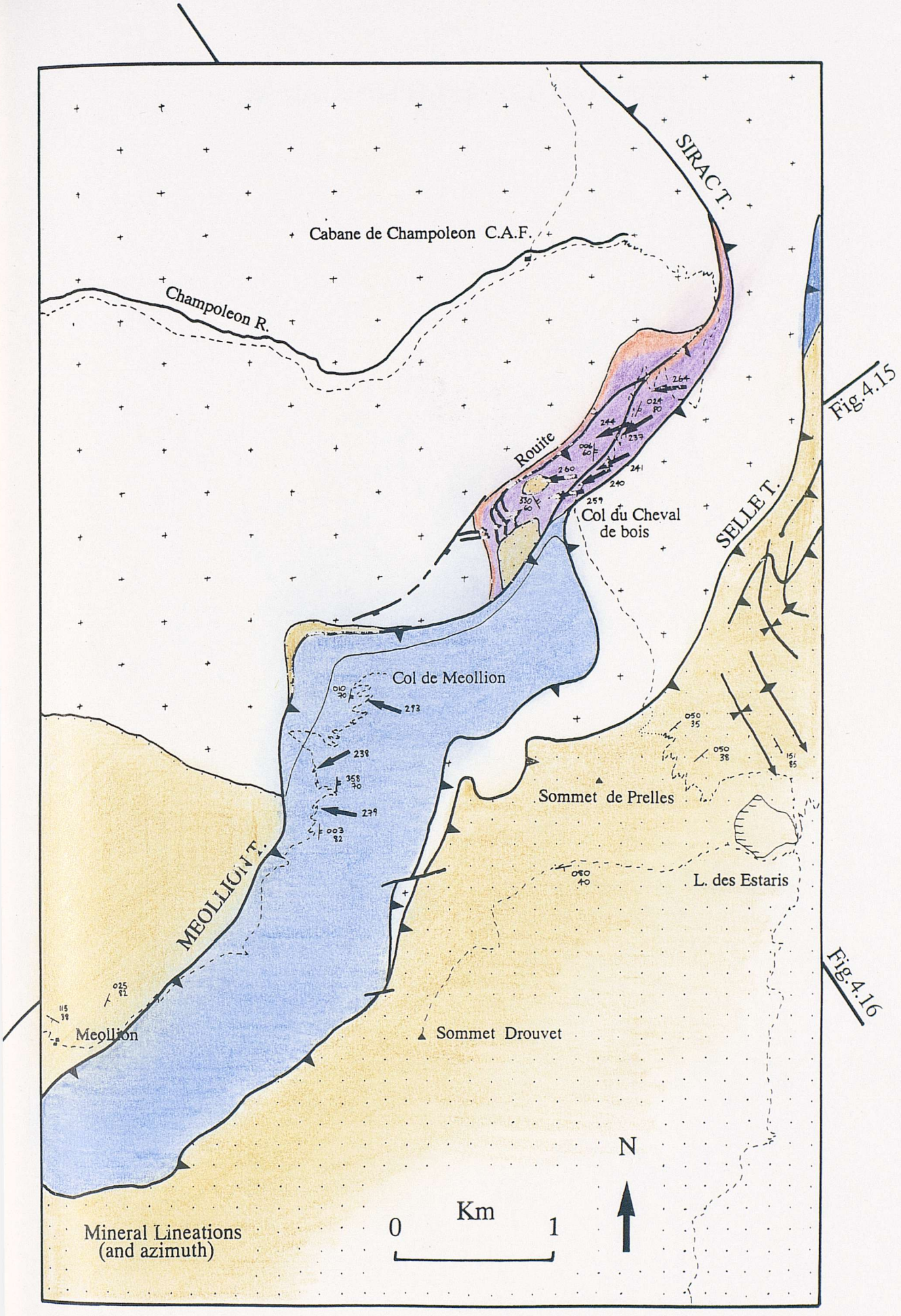
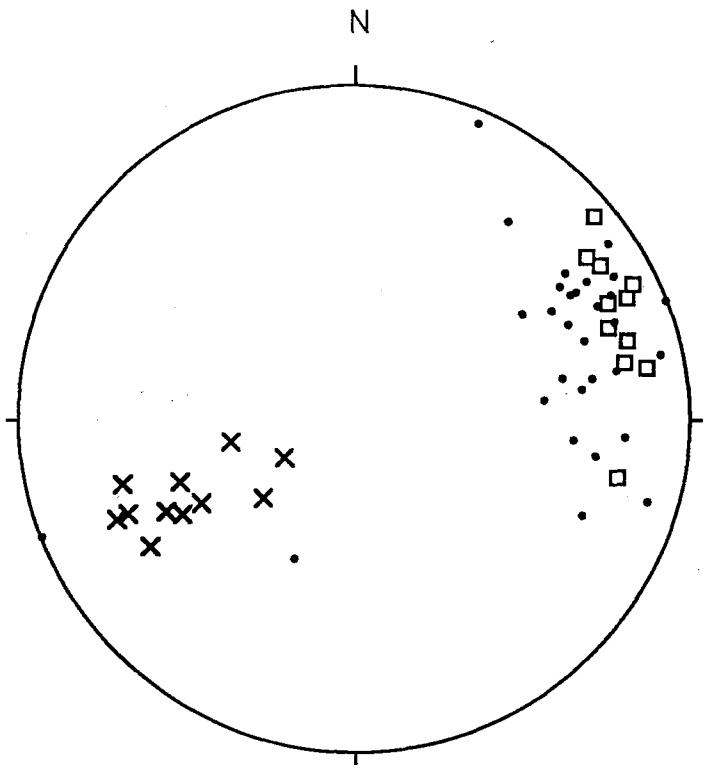


Fig.4.14. MAP OF THE MEOLLION–ROUGNOUX LOCALITY, SOUTHEASTERN PELVOUX,
showing transport direction indicators in the footwall to the Sirac Thrust.

Mineral Lineation Data from SE. Pelvoux



Key

- • 29 Data from Meollion / Rougnoux
- x 11 Cleavage Poles from Selle
- □ 11 Data from Dourmillouse

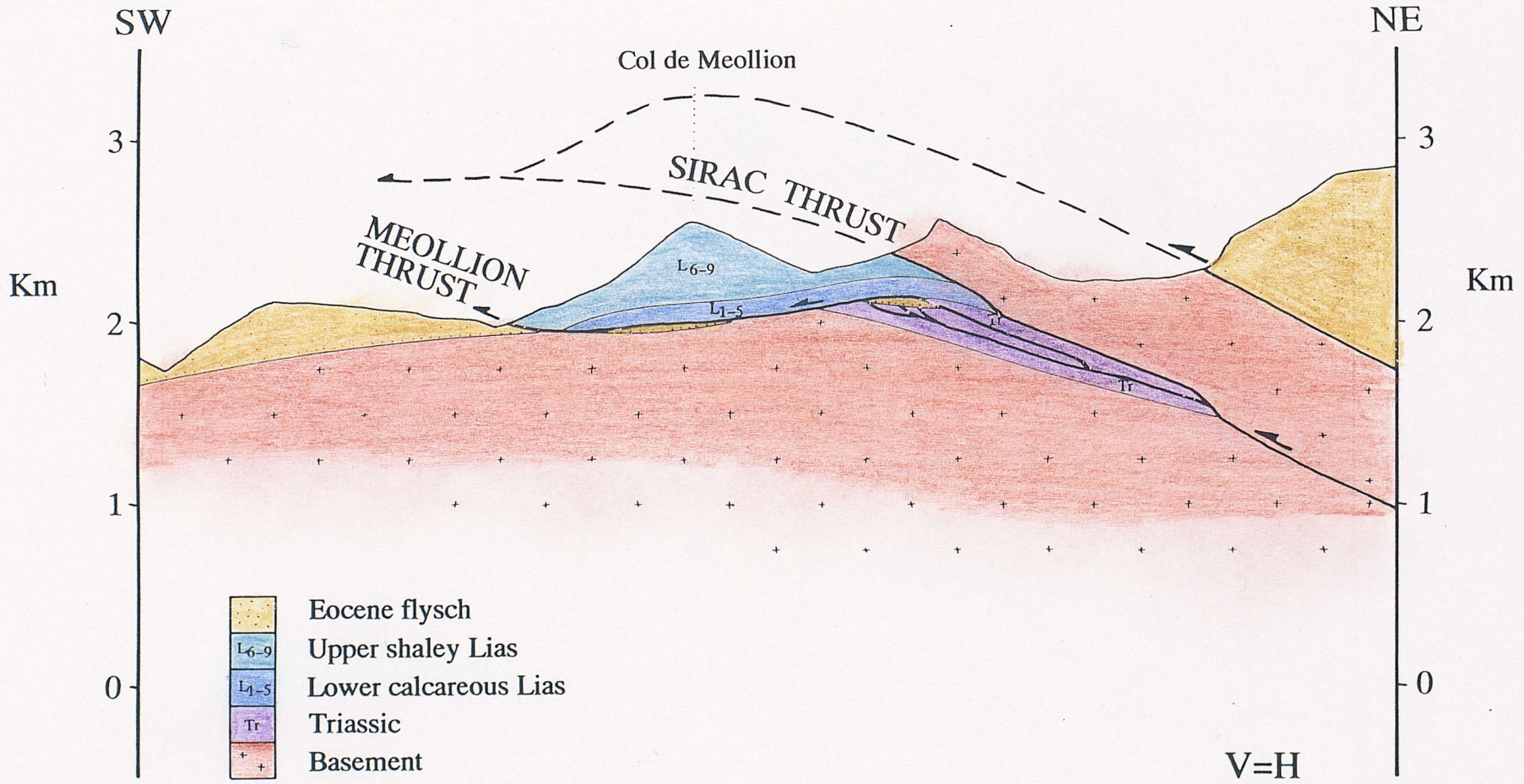


Fig.4.15. SECTION ACROSS THE MEOLLION–ROUGNOUX AREA,
drawn parallel to the post–Nummulitic transport direction.

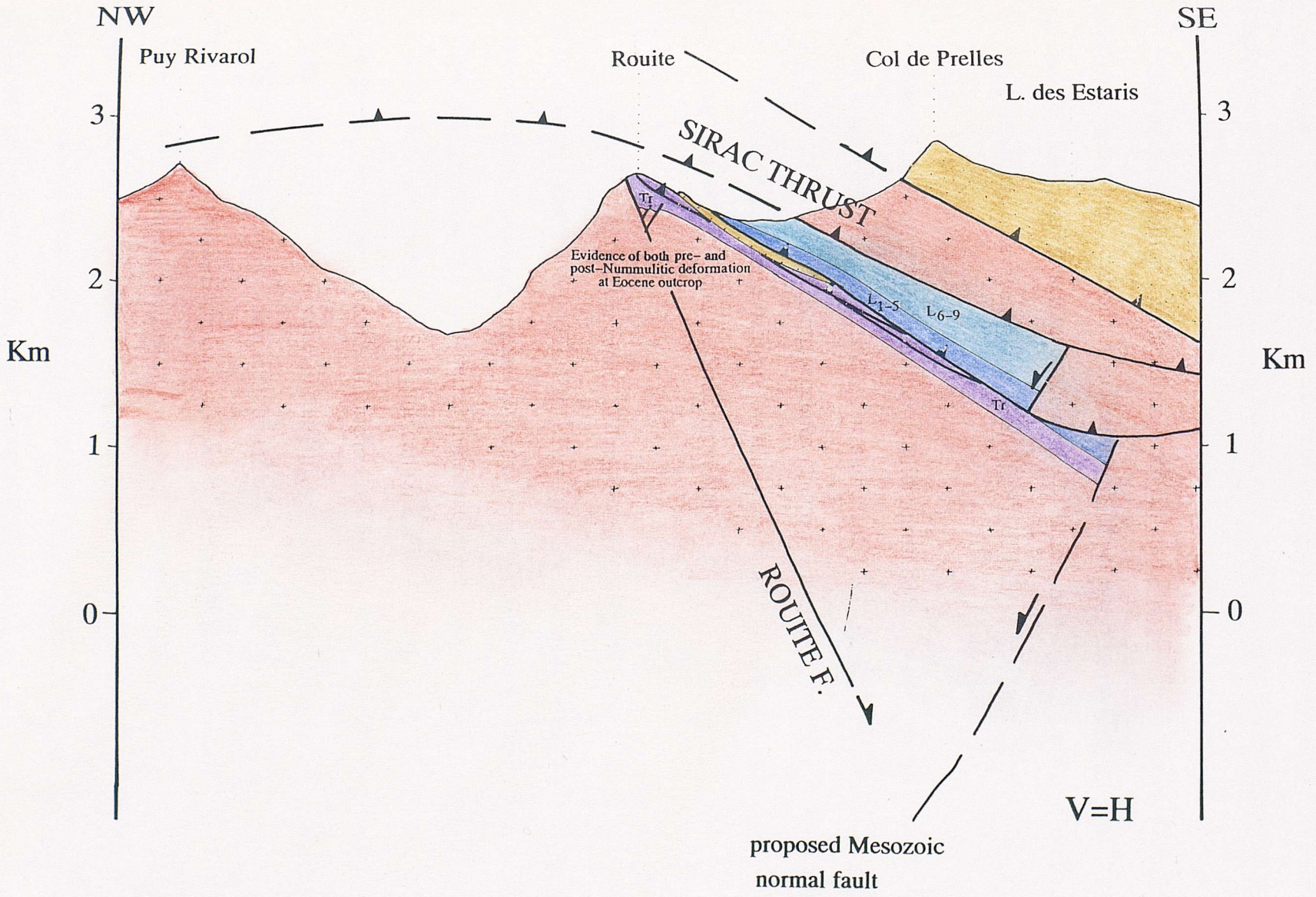


Fig.4.16. SECTION ACROSS THE MEOLLION–ROUGNOUX AREA,
drawn normal to the post–Nummulitic transport direction to demonstrate the Route extensional fault geometry. Note the proposed graben–bounding fault to the east.

The southern flank of the Aiguille de Cedera is a dip slope composed of Eocene flysch sediments that strike ENE (N075°–082°E) and dip at 36–38°S. Locally the dip reaches 80°S due to local strains. The post–Nummulitic tilting of the region is therefore greater at this locality than in either Tourond or Muande where the Eocene unconformity dips at approximately 30°SSE.

East of the Méollion farmstead the Domerian and Toarcian sequence is exposed. The micaceous marls, shaley carbonates and calcareous shales of the upper Lias have been thrust over the Gres du Champsaur flysch sequence on the Méollion thrust. The hangingwall sediments are strongly foliated, the cleavage striking N–S (N358°–010°E) and dipping steeply at 70–82°E.

North of the Col de Méollion, the Méollion thrust slice is well–exposed; lower Lias carbonates being juxtaposed against a thin slither of basal Eocene conglomeratic limestone (Figs.4.14 & 4.15). The overlying sequence of upper Lias shales has been weathered away to form the col itself, and these beds are overthrust by the Sirac basement slab. Pods of Triassic spilite have been sheared along the Sirac thrust plane.

In the Rougnoux valley a veneer of Triassic strata is trapped between the Cedera/Rouite and the Sirac basement sheets (Fig.4.15). This unit has been imbricated and strongly sheared, producing mineral lineations directed towards the west–southwest (N240°–261°E). At several outcrops strongly sheared spilites occur immediately below coarse Eocene conglomerate. The age of this shearing is uncertain. Since the conglomerates manifest no evidence of shear the spilite lineations may be pre–Eocene. Imbricates of spilite to the southwest of Col du Cheval de Bois are truncated by the Eocene unconformity indicating that pre–Eocene deformation did occur. However, since major movement on the Sirac basement thrust was post–Eocene it is certain that lineations of post–Eocene age also exist. During this phase the coarse Eocene conglomerates may have behaved as resistant blocks within a deforming matrix of Triassic spilite. It appears therefore that two phases of shearing have produced the mineral lineations within the spilite. The phases were approximately colinear since the data has a small spread (N240°–261°E), but the post–Eocene phase was the more intense.



a) VIEW OF AIGUILLE DE CEDERA FROM THE SOUTH, showing steeply-dipping "Gres de Champsaur" flysch unconformably overlying basement.



b) VIEW OF SHEARED TRIASSIC SPILITE AT COL DU CHEVAL DE BOIS, with unconformable Eocene conglomerate to the right. Lineation to WSW.



VIEW OF COL DE MEOLLION FROM COL DU CHEVAL BE BOIS,
showing the post–Nummulitic Sirac and Meollion Thrusts.

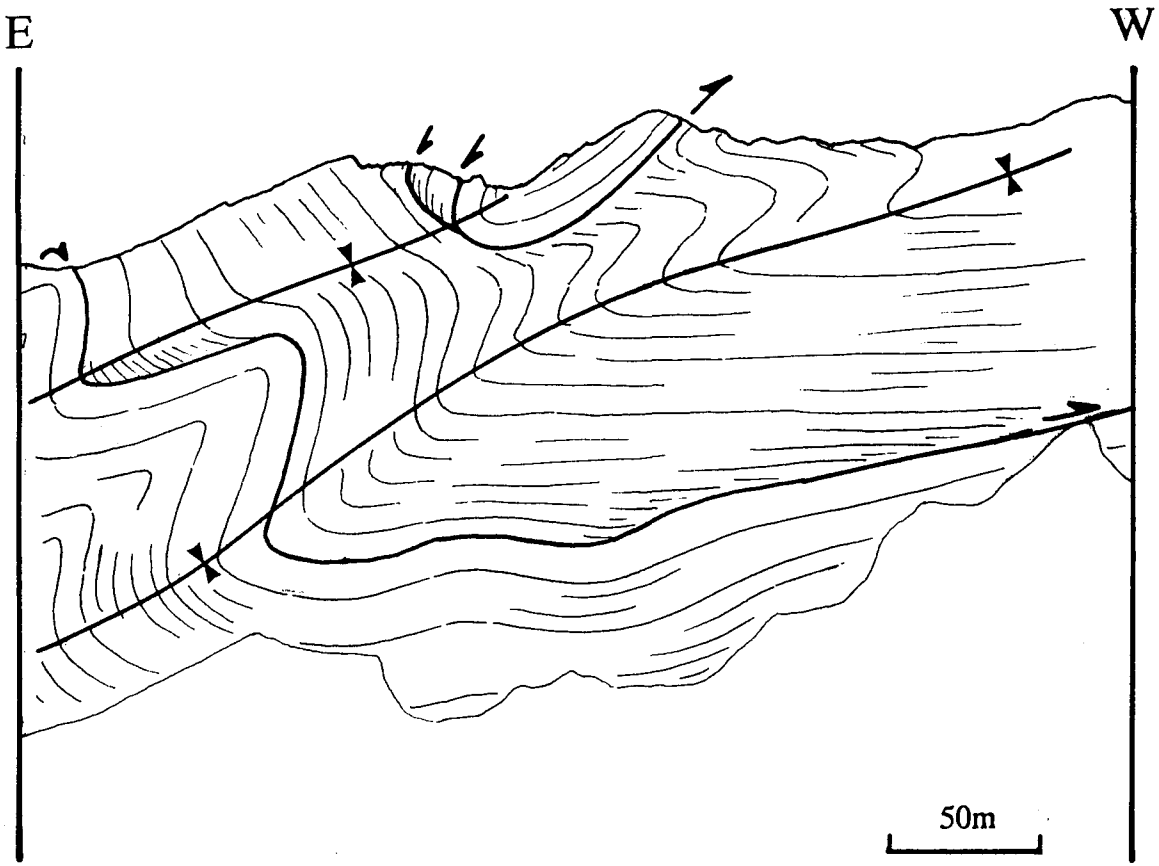
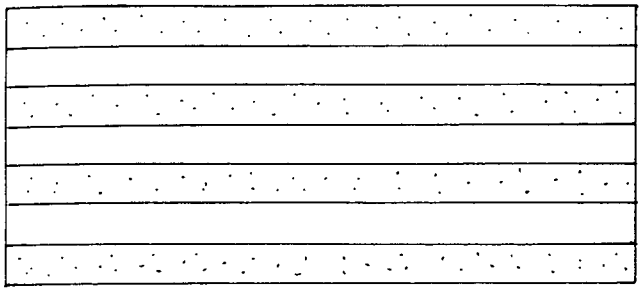
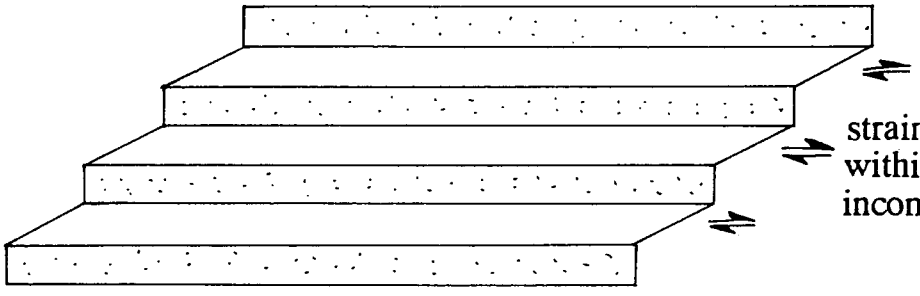


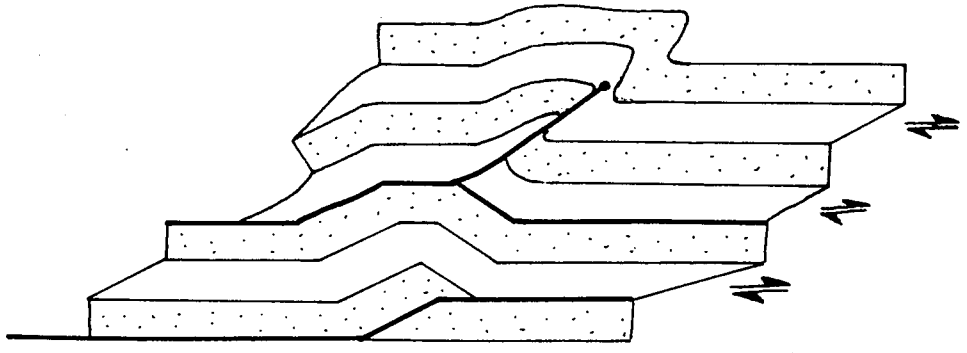
Fig.4.17. FIELD SKETCH OF THE EOCENE FLYSCH EXPOSURE AT CRETE DU MARTINET, ROUGNOUX.



1) ALTERNATING SERIES OF COMPETENT AND INCOMPETENT UNITS.



2) IDEALISED LAYER-PARALLEL SHEARING (LPS),



3) ACTUAL LPS, shears ramp from detachment to detachment and terminate as tipfolds.

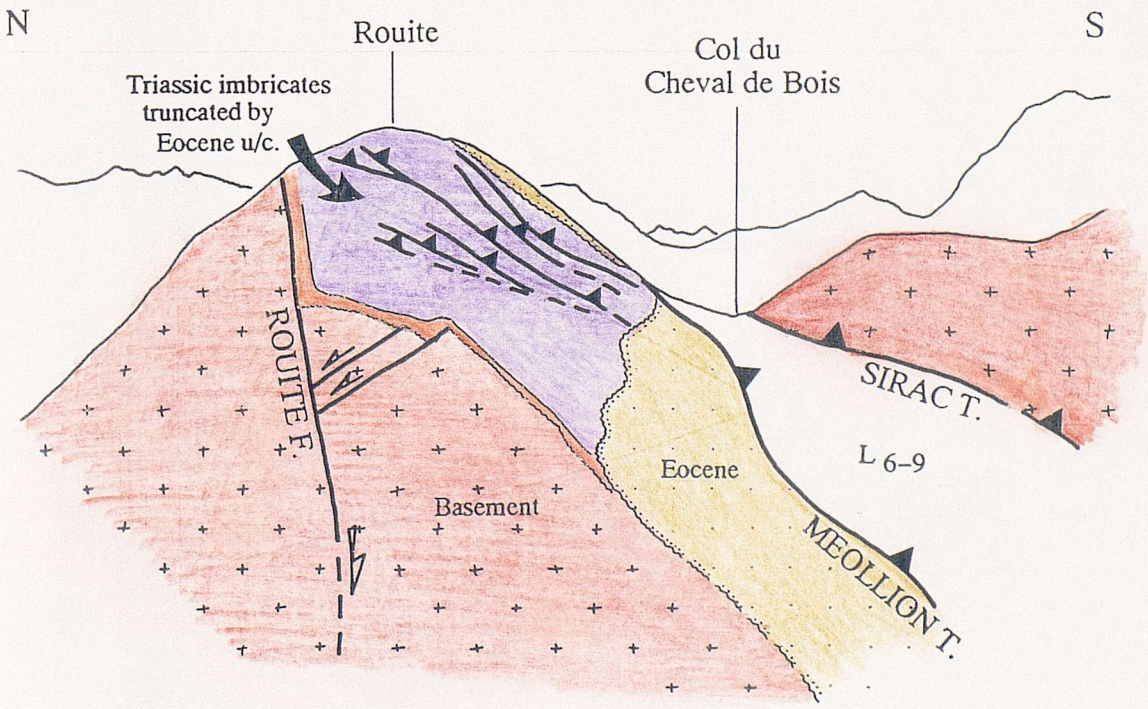
Fig.4.18. CARTOON ILLUSTRATING THE DEFORMATION STYLE WITHIN THE FLYSCH SEQUENCE.

The Eocene flysch series has been folded in the hangingwall of the Selle thrust. The basal Eocene units, if present, are concealed by scree in the Rougnoux valley and only the Gres du Champsaur unit is exposed. The flysch sands are spectacularly folded at Rougnoux (Fig.4.17). Vertical beds strike approximately N170°–180°E indicating W–vergent deformation. These folds can also be observed from the Chichin valley to the south.

The Rougnoux flysch outcrop exhibits thrusts that have been folded during progressive post–Eocene shearing. This is only revealed by footwall and hangingwall ramp cut–offs since thrust flats are inconspicuous in the monotonous sand–shale series. This characterises progressive deformation where numerous parallel, easy–slip horizons, i.e. the shale members of the flysch succession, alternate with more competent layers, i.e. the flysch sandstones. In the absence of ramping, the flysch sequence would deform like a "deck of cards". In reality, faults splay from the layer–parallel detachments and either breach or fold the overlying strata and any structures already developed within it (Fig.4.18).

Important Mesozoic extensional faults are exposed to the west of the Col du Cheval de Bois (Figs.4.14 & 4.16). From the Col de Méollion the basement and Triassic dolomites are clearly offset. The Triassic spilites form drape folds across the fault offsets. The faults trend NE–SW and comprise a major fault downthrowing to the southeast and two minor antithetic faults which have a downthrow of 5–8m. The displacement on the major fault is unknown due to erosion, but the minimum throw is approximately 140m. This fault is part of a set of NE–SW trending extensional faults but is thought to be of relatively minor significance – being an antithetic to a proposed half–graben bounding fault to the southeast. Although this proposed fault does not outcrop its presence is deduced from the Eocene subcrop pattern; Mesozoic sediments are present in the Selle valley, but absent at Fournel, and only thin Triassic strata are preserved at Dourmillouse. This suggests the presence of a NE–SW trending fault, downthrowing to the northwest, such that the Selle sequence represents half–graben fill whilst the Fournel and Dourmillouse outcrops correspond to the uplifted footwall shoulder of a basement fault block. The proposed fault lies between the Selle and Fournel localities (Fig.4.19). This feature will be dis–





VIEW OF ROUITE AND THE COL DU CHEVAL DE BOIS FROM THE EAST, showing Mesozoic extensional faults and pre– and post–Nummulitic Alpine thrusting

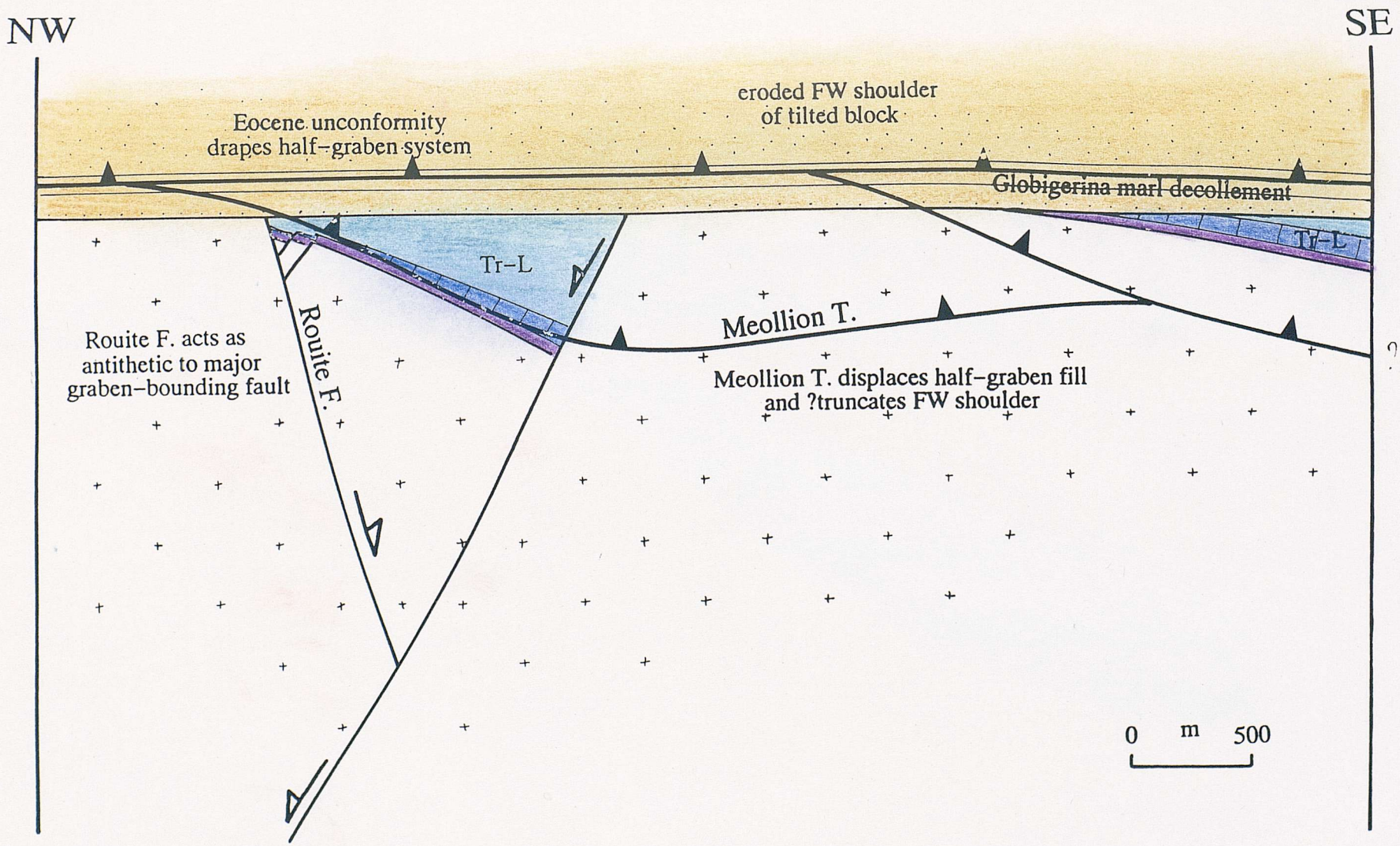


Fig.4.19. CARTOON ILLUSTRATING THE PROPOSED TILTED FAULT BLOCK GEOMETRY PRIOR TO POST-NUMMULITIC THRUSTING.

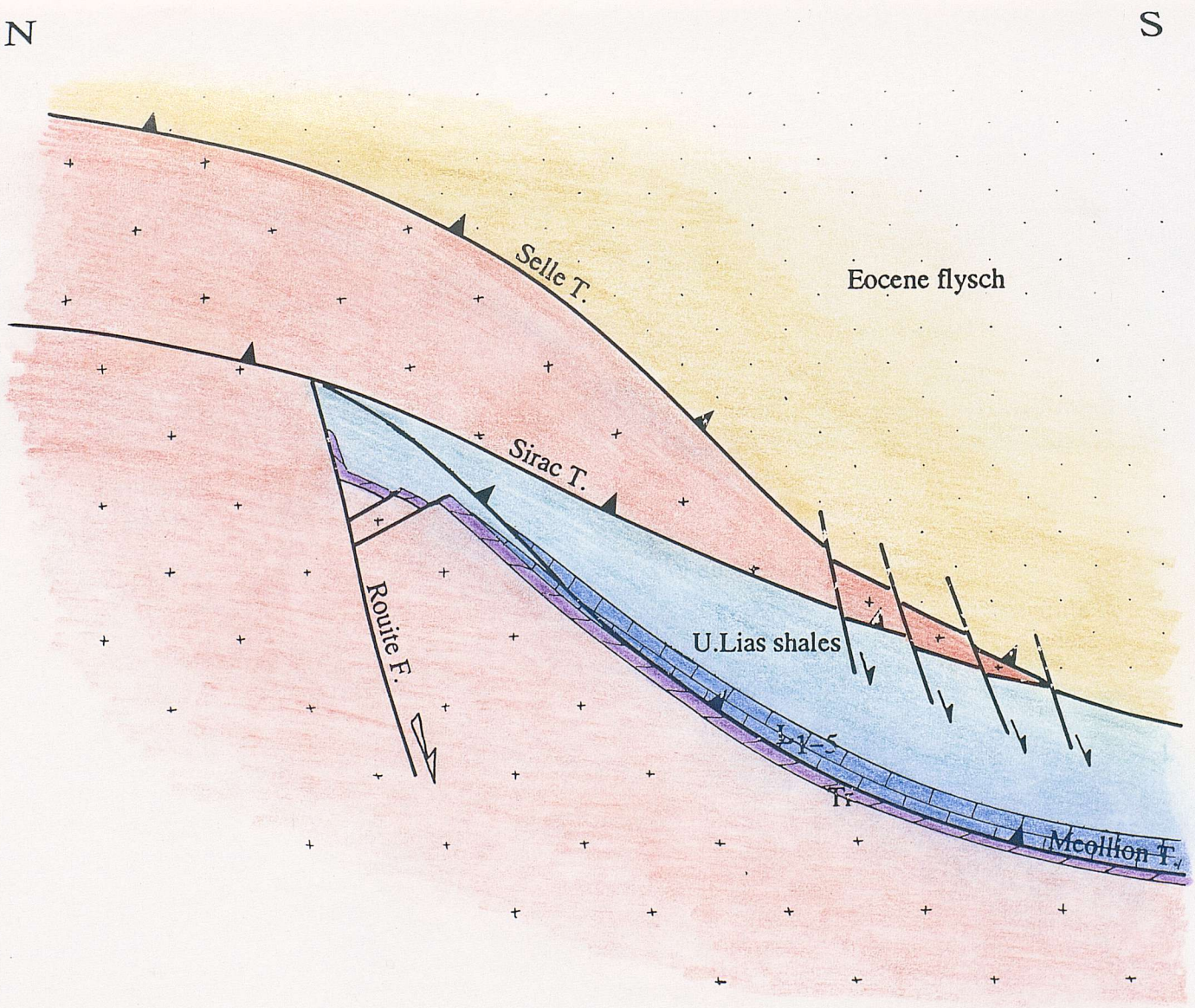
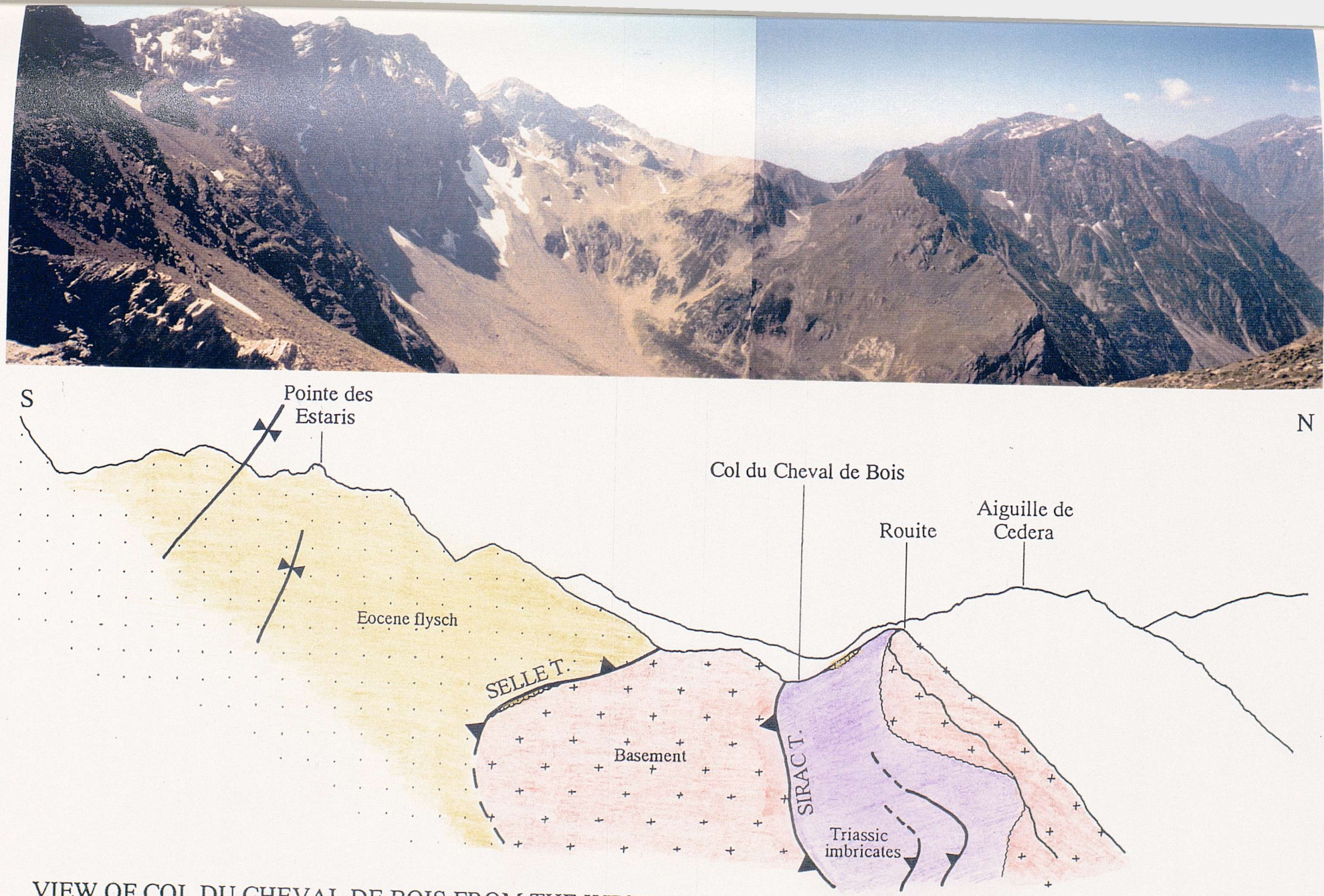


Fig.4.20. SKETCH SECTION THROUGH LATERAL HW RAMP OF SIRAC THRUST,
showing proposed collapse of flysch overburden into Liassic shales of the Meollion thrust sheet



VIEW OF COL DU CHEVAL DE BOIS FROM THE WEST,
showing folded Eocene flysch in the Crete de Martinet (left) and Route basement and cover in the footwall of the Sirac basement thrust.

cussed more fully later (see section 4.2.4).

From Pnte. des Pisses and Sommet Drouvet (2655m) late extensional faults are seen to offset the Sirac basement thrust and the overlying flysch (Fig.4.14). These faults probably developed during movement on the Méollion thrust. The Sirac and Selle thrust sheets were carried passively in the hangingwall of the Méollion thrust. Because of the weakness of the Méollion sheet, mainly comprising the upper shaley Lias sequence, the overburden of the Sirac basement and Selle flysch packages collapsed into the underlying shales (Fig.4.20). In most places the Sirac basement sheet was too thick to collapse in this manner. However, in the hangingwall ramp of the Sirac thrust where the basement is thinner, the basement cannot support the thick flysch series and collapses into the underlying shales. The normal faults trend approximately E–W and downthrow predominantly to the south by 5–25m. Their trend can be related to the orientation of the Sirac thrust ramps (c.f. La Muande locality, section 4.1.2).

4.2.2. Selle

The Selle valley lies just northeast of the Rougnoux valley, being connected via the Pas de la Cavale (2735m) and the Col de l'Aup Martin (2761m). Vehicular access is, however, from the east via l'Argentiere–la Bessée, Vallouise and Entre les Aigues, from where one can ascend on foot to the Col de l'Aup Martin.

The outcrops of interest lie at the head of the valley, between the Col de l'Aup Martin and the Pointe de Neyzets on the northwestern flank of the Crete de l'Aup Martin (Figs.4.21 & 4.22). Examples of both pre– and post–Nummulitic deformation are exposed in the Hercynian basement and Mesozoic/lower Tertiary cover sediments. The basic structure is a large pre–Nummulitic basement syncline, similar to the Muande and Cedera folds, that has suffered post–Nummulitic modification. As one moves eastwards from the St. Bonnet to the Selle outcrops, the intensity of the post–Nummulitic deformation becomes noticeably greater. Whereas in St. Bonnet only the Nummulitic flysch series is affected by post–Nummulitic shearing, in Rougnoux and Selle the basement and Mesozoic strata are also involved. Hence, in the west the phases affected different

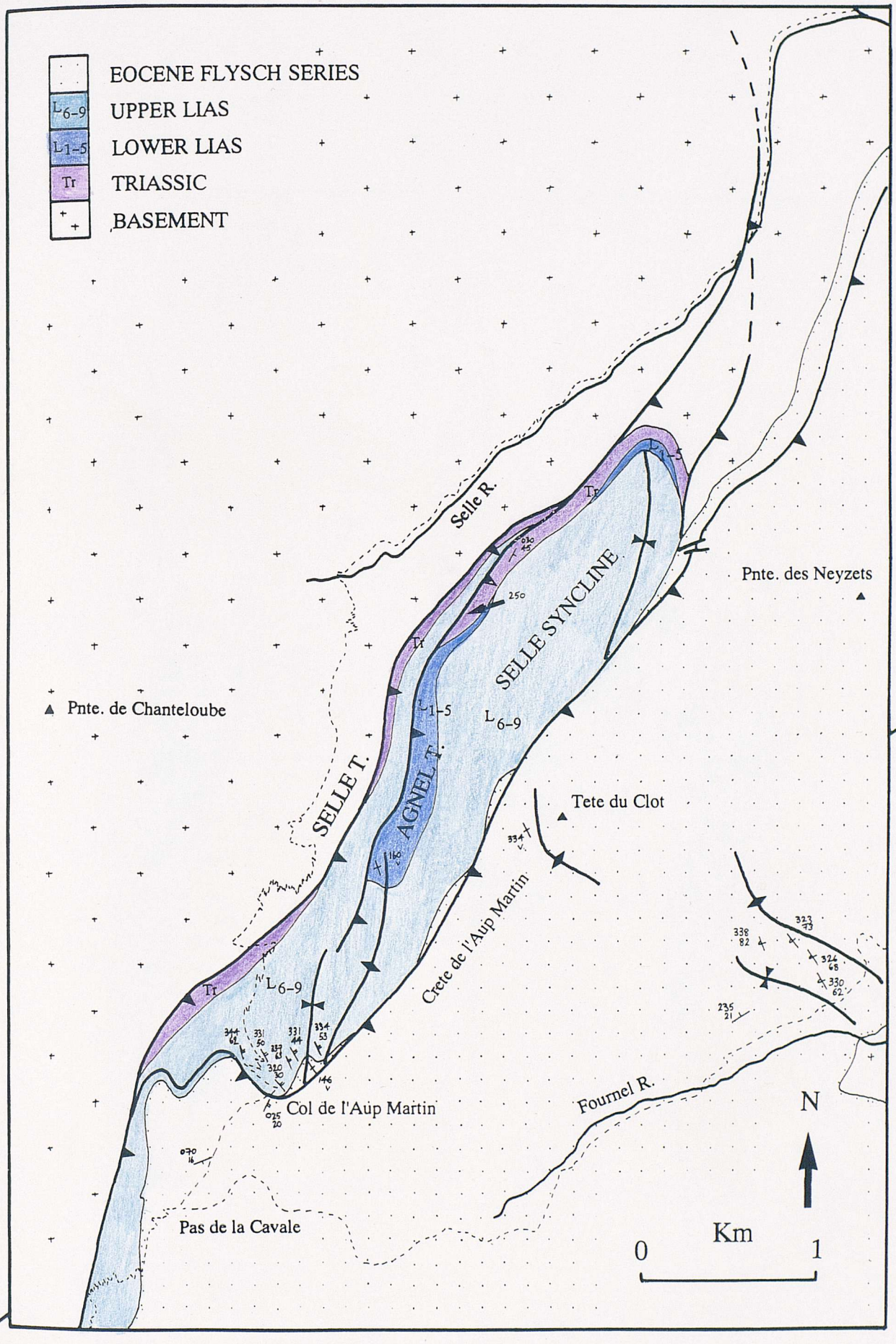


Fig.4.21 MAP OF THE SELLE LOCALITY, SOUTHEASTERN PELVOUX

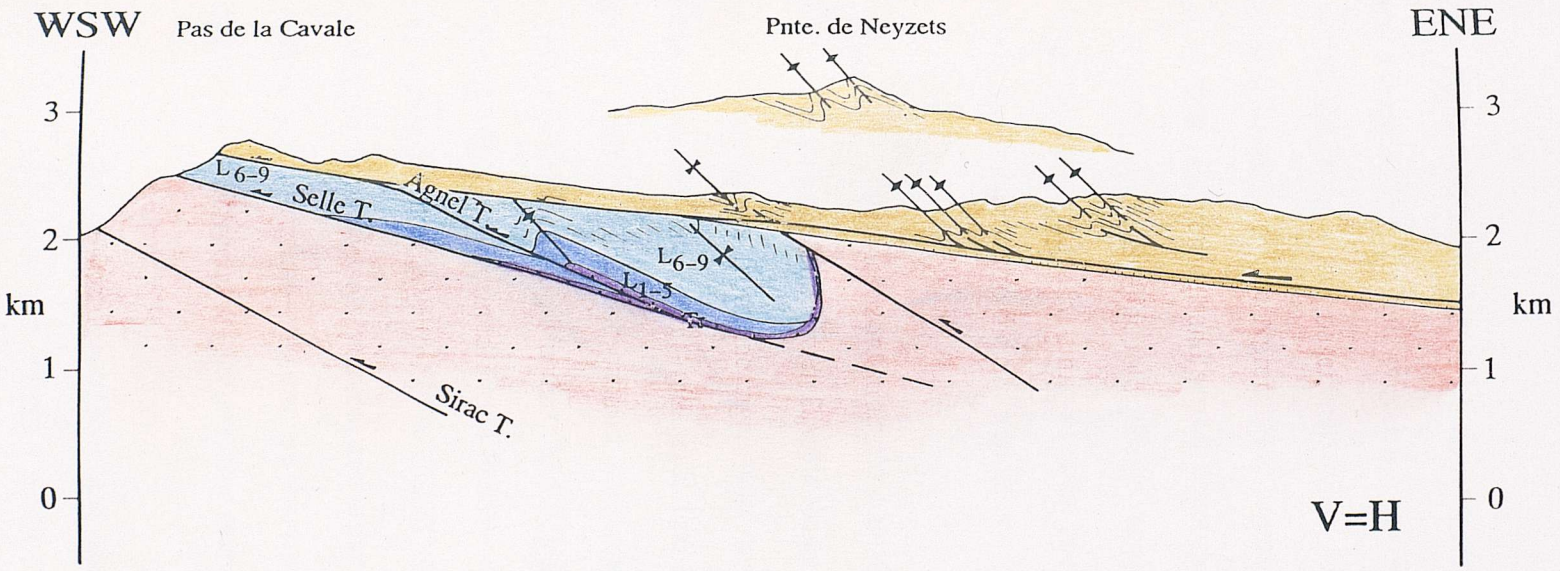
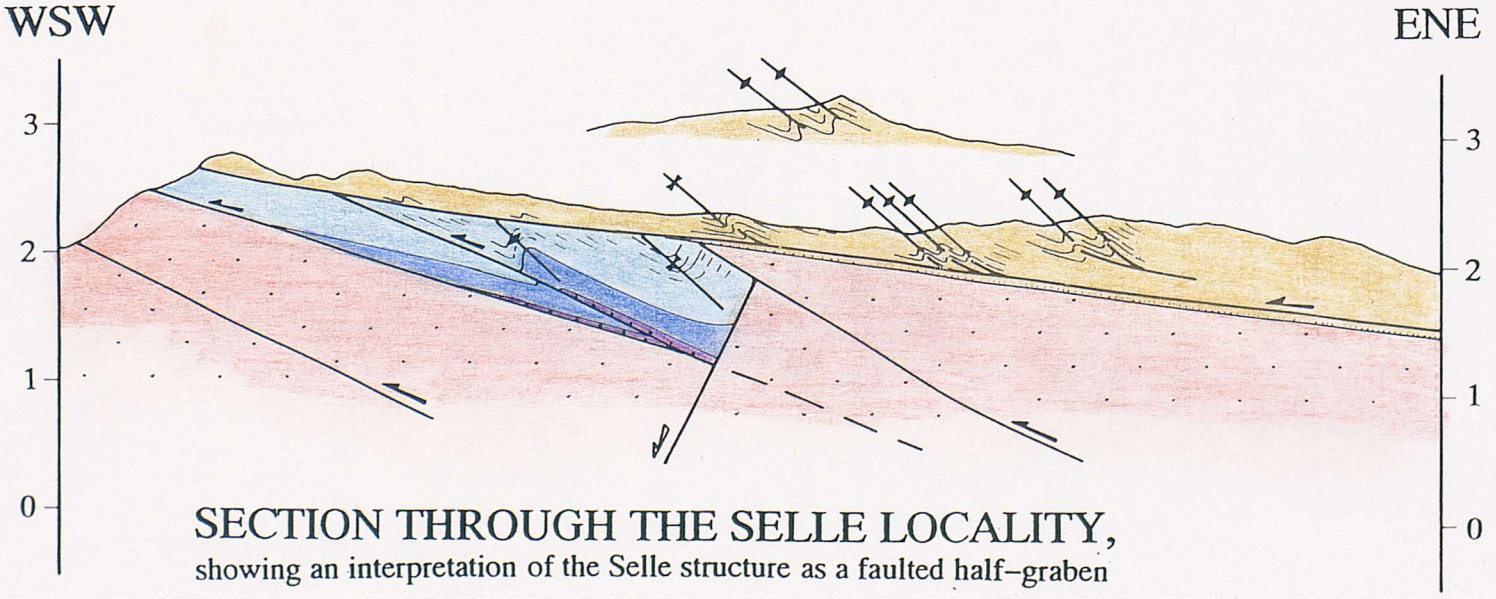


FIG.4.22 SECTION THROUGH THE SELLE STRUCTURE,
showing modification of the pre–Eocene Selle syncline by post–Eocene shearing



SECTION THROUGH THE SELLE LOCALITY,
showing an interpretation of the Selle structure as a faulted half–graben

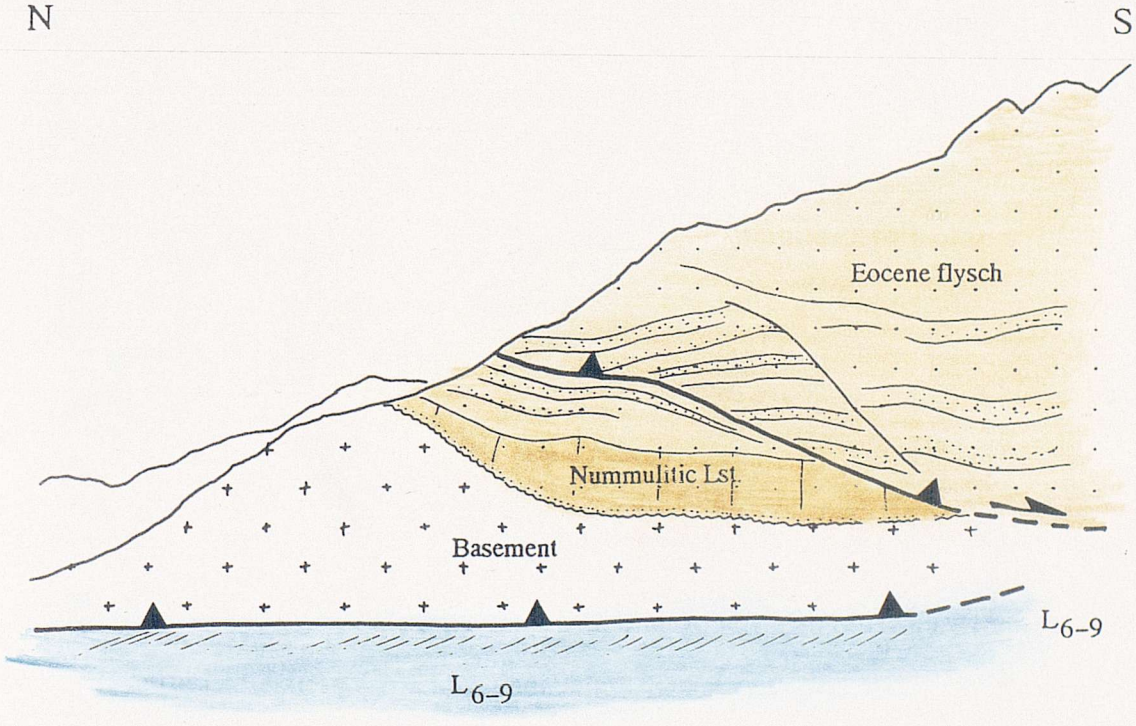
structural levels and are therefore quite distinct, whilst in the east the Mesozoic strata have suffered from the superposition of both pre- and post-Nummulitic phases.

Northwest of Pointe de Neyzets the Hercynian basement and the Mesozoic succession form a large WSW-vergent syncline - **the Selle syncline**. The overturned limb of the fold has been breached by a late basement thrust. This structure is of pre-Eocene age, being unconformably overlain by the Nummulitic limestone. The Selle fold is assymmetric; the overturned fold limb is sub-vertical and strikes towards N150°-160°E whilst the right-way-up limb dips gently to the southeast. The fold plunges at approximately 10°SE. The regional post-Nummulitic tilting therefore gradually diminishes as one moves eastward from Muande, through Tourond, Cedera and Rougnoux to the Selle locality.

The Mesozoic strata do not form a coherent outcrop pattern around the Selle fold. Rather, the Triassic and lower Liassic strata are impersistent on the overturned fold limb. The Agnel thrust splays from the right-way-up limb of the Selle fold. The Liassic strata in its breached tipfold have a cut-off angle that exceeds 90°, the vertical lower Liassic limestones striking N160°E. The unbreached continuation of this tipfold outcrops two hundred metres northeast of Col de l'Aup Martin (Fig.4.21). Here the Nummulitic limestone and the upper shaley Lias sequence are deformed into a WSW-vergent fold pair with a strong axial planar cleavage. The cleavage strikes NW to NNW (N320°-344°E) and dips moderately at 30-62°E. The overturned fold limbs are sub-vertical and strike at N146°-155°E degrees. The Agnel thrust formed by pre-Nummulitic out-of-syncline thrusting but the involvement of the Nummulitic limestone means that it suffered post-Nummulitic reactivation.

Post-Nummulitic deformation has strongly overprinted and modified the Selle syncline (Fig.4.21) which now lies within a network of post-Nummulitic shears. The Selle and Agnel thrusts both flatten at Eocene levels. Both these thrusts rise into a roofthrust in the Eocene Globigerina marl unit. This roofthrust cuts down-section in places, plucking pods of the basal Eocene units from its footwall and the breaching the Eocene unconformity to incorporate Liassic shales





VIEW OF THE NORTHEASTERN PART OF THE SELLE SYNCLINE, showing the Nummulitic limestone lying unconformably upon the basement and post–Nummulitic shearing cutting down from the flysch to the Liassic shales.



VIEW OF POST-NUMMULITIC FOLDS AT COL DE L'AUP MARTIN, showing folds in the Upper Lias shales and Nummulitic Lst. being truncated by the Eocene flysch.

below the Crete de l'Aup Martin. Hence, the Nummulitic unconformity is strongly sheared for much of its outcrop in the Selle valley.

It is significant that the Eocene unconformity is sheared only where it overlies the shaley Liassic facies, whilst where it overlies crystalline basement the shearing remains within the Globigerina marl unit. Hence below Pointe de Neyzets the unconformity remains intact but at Col de l'Aup Martin it is intensely sheared.

4.2.3. Fournel

The Fournel river flows east out of the Pelvoux massif to join the Durance river at l'Argentiere-La Bessée. Ascending the valley, one passes through the Brianconnais Zone and the Frontal Pennine Thrust before entering the external Dauphinois Zone. Vehicular access is possible as far as Haute Salce, from where a footpath leads to Pas de la Cavale (2735m).

Unlike the other localities described above, at Fournel no Mesozoic strata outcrop (Fig.4.23). Instead the Hercynian basement is directly overlain by the Nummulitic trilogy. Hence post-Nummulitic deformation is most clearly observed in the field; the flysch series being sheared and folded in spectacular fashion. Evidence of pre-Nummulitic deformation is more cryptic, but equally important. This early Alpine phase produced strong uplift prior to the deposition of the flysch series in late Eocene times, resulting in the exposure of basement rocks on the Priabonian landscape.

Within the scope of the Fournel outcrop, the pre-Nummulitic topography was relatively planar. However the basal conglomerate and Nummulitic limestone member of the Nummulitic trilogy does exhibit abrupt thickness variations indicating the presence of minor fault scarps. Two such faults are mapped at the Fournel window. They strike NNW-NW and downthrow to both the east and west by 5-7m. The Globigerina marl unit is flat-bedded and unaffected by these offsets.

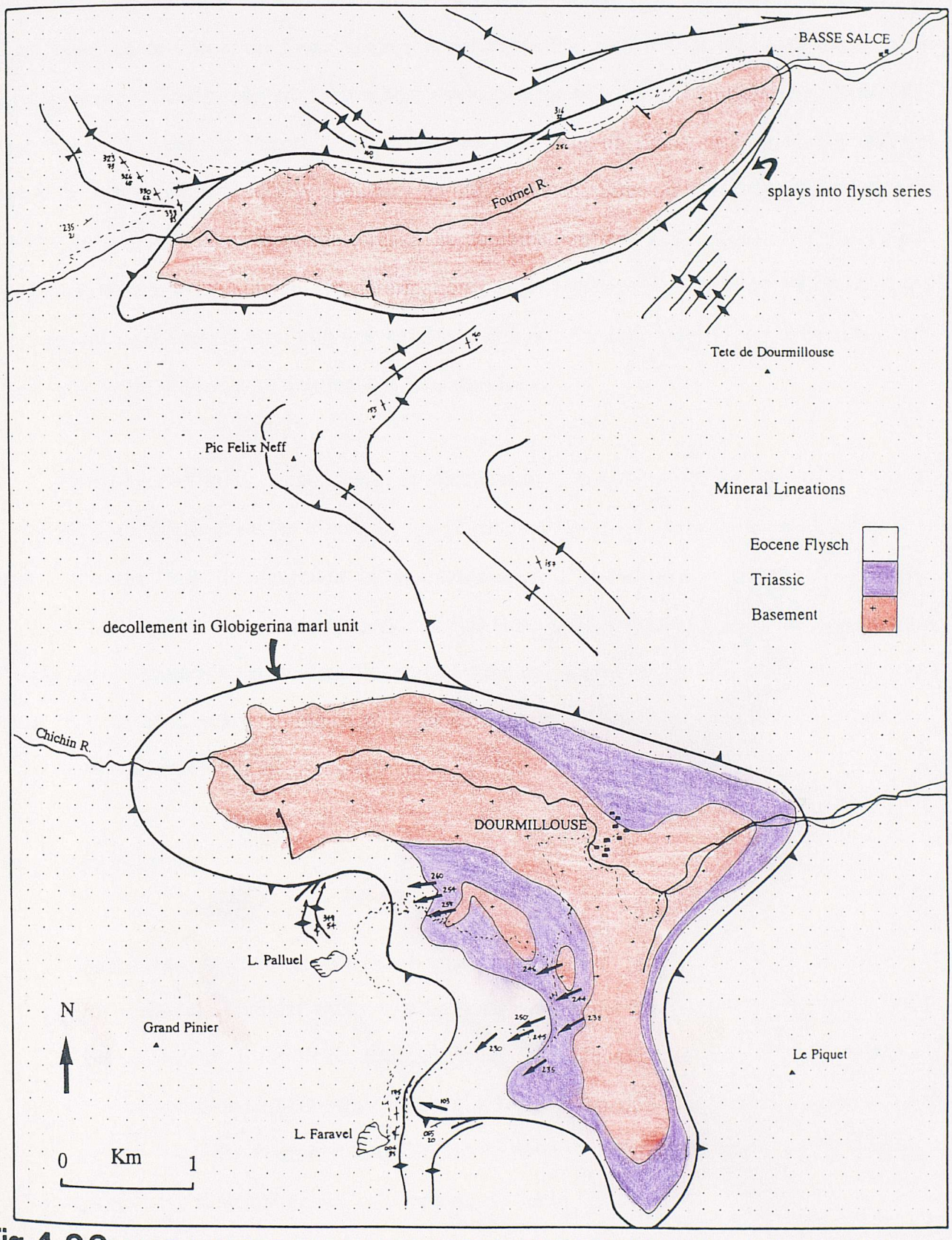


Fig.4.23 MAP OF THE FOURNEL AND DOURMILLOUSE LOCALITIES, SOUTHEASTERN PELVOUX.

Note the absence of Mesozoic strata within the area of the map, except in southeastern Dourmillouse where a thin veneer of Triassic strata outcrop beneath the Nummulitic unconformity. This indicates pre–Nummulitic uplift related to either Mesozoic rifting or early Alpine thrusting. The region is dominated by W–WSW vergent post–Nummulitic shearing.

Mineral lineation data from the Globigerina marl unit indicates WSW-vergent shear (N256°E). This is supported by cleavage foliations that strike NW-SE (N316°E) and dip at 32°E. Bedding data from the flysch sediments define the fold orientations. The beds form a train of assymmetric, WSW-vergent folds whose overturned or sub-vertical limbs strike NNW-NW (N323°-340°E) and dip steeply at 62-90°E. The right-way-up limbs strike SW-NE and dip gently at approximately 20°NW, contrary to the consistent SSE-directed tilt previously described in the footwall of the Sirac thrust. The combined structural data provides an excellent constraint on the post-Nummulitic deformation phase. During this period of WSW-directed shearing the Globigerina marl unit was exploited as a decollement, from which splays rose into the overlying flysch series to terminate as large tip-folds.

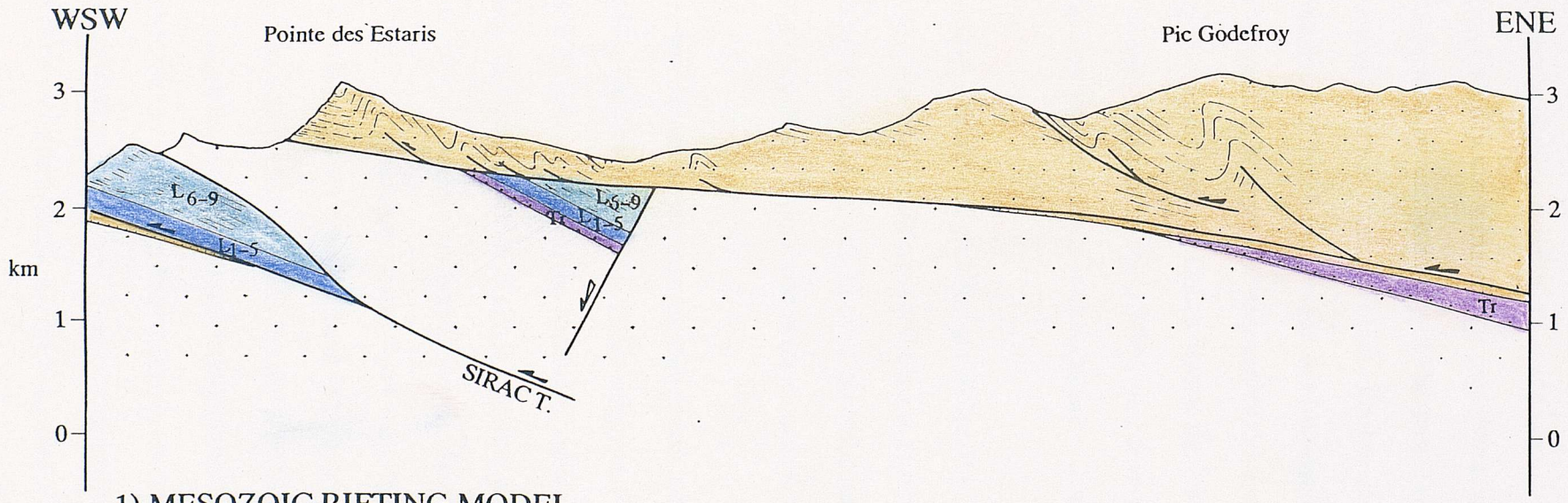
As mentioned above, small-scale pre-Nummulitic deformation is not observed at Fournel. However, on a larger scale the effects of this phase, and the pre-existing Mesozoic extensional phase, are clear from the basement-cover relationships. The presence or absence of Mesozoic sediments in southern Pelvoux is a function of two factors, 1) the Mesozoic palaeogeography and 2) the pre-Nummulitic uplift. The Priabonian subcrop pattern results from a combination of early Jurassic extensional faulting and Palaeocene-early Eocene basement overthrusting. The details of this subcrop will be discussed later.

4.2.4. Dourmillouse

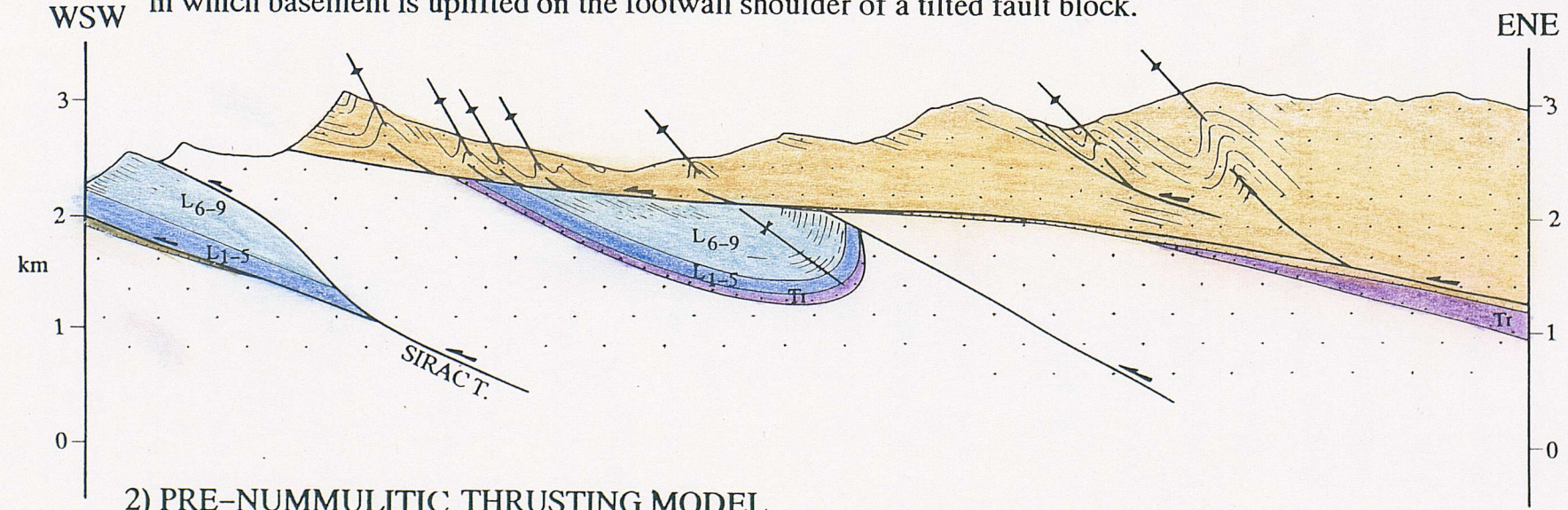
Dourmillouse is a village that lies at approximately 1700m in the Chichin or Biaysse valley, which joins the Durance valley near La Roche-de-Rame. Dourmillouse is approached by road via Fressinieres, but the final ascent is by foot. The village is surrounded by high ridges. The Tete de Dourmillouse (3084m) and Pic Felix Neff (3242m) lie to the north, separating the Biaysse valley from the Fournel valley. To the southwest lies Grand Pinier (3117m), on the slopes of which the principal geological outcrops are found.

At the Dourmillouse window (Fig.4.23), Hercynian basement and Triassic strata are

Fig.4.24. TWO SECTIONS ACROSS THE DOURMILLOUSE–MEOLLION STRUCTURE,
showing post–Nummulitic shearing in the flysch and two possible causes of pre–Nummulitic uplift that juxtaposed basement against the Nummulitic unconformity. For location see Fig.4.13



1) MESOZOIC RIFTING MODEL,
in which basement is uplifted on the footwall shoulder of a tilted fault block.



2) PRE–NUMMULITIC THRUSTING MODEL,
in which basement is uplfted in the hangingwall of a "pli–faille" structure (cf. Cedera).

exposed through a thick blanket of unconformable Eocene–Oligocene flysch. To the east of Dourmillouse the flysch sediments are exposed at road level, but towards the west the Eocene unconformity slowly rises indicating post–Nummulitic uplift (Fig.4.24). An earlier period of tilting is indicated by eastward dip of the base Triassic unconformity (Figs.2.23 & 2.24). The Triassic unconformity is truncated by the Eocene unconformity; juxtaposing Hercynian basement against Nummulitic flysch in the west, but introducing Triassic dolomites and spilites in the east. As noted at Fournel, Nummulitic–age extension occurred prior to the onset of flysch sedimentation. Two kilometres west of Dourmillouse the basal Eocene conglomeratic carbonates show a sharp thickness variation across a NNW–trending fault that downthrows to the west indicating synsedimentary faulting. The flysch sequence is unaffected.

The main Alpine deformation at this locality is post–Nummulitic, but the inability to distinguish pre–Nummulitic deformation does not preclude its presence. No mapped compressional structures are truncated by the Eocene unconformity and shears within the Triassic spilites are statistically inseparable from those in the Eocene Globigerina marl unit.

Mineral lineations are strongly developed in the Triassic spilite formation, near the path leading to L. Faravel (Fig.4.23). Their azimuths trend approximately WSW, varying between N235°–246°E. Near L. Palluel the lineations are more E–W oriented (N258°E). This variation is well within the data scatter of a single shearing phase and two phases of deformation are not postulated. Lineation azimuths within the Globigerina marl unit mimic those from the Triassic spilites (N230°–250°E in the south, N254°–260°E in the north). Bedding data from the flysch succession indicates west–vergent shear. The large–scale folds are overturned towards the west, the overturned limbs striking N–S (N168°–186°E) and dipping at 54–90°E. The data suggests therefore that the transport direction swings from WSW to W as one rises up–section from near the basement–cover contact to within the flysch basin. However the clockwise statistical swing in transport direction is only 3–4° between the spilite and the marl, and 13° between the marl and the flysch. It seems more likely that the variations are local anomalies reflecting basement topography and that the prevailing transport direction was towards N255°–260°E.

Large folds are observed in the flysch on the southern flank of Pic Felix Neff. Vertical beds within these outcrops strike at N157°E, correlating with beds at Fournel that strike N155°–160°E, and suggesting tectonic transport towards the WSW.

In section 4.2.3. the configuration of the basement and Mesozoic cover beneath the Eocene unconformity was mentioned. At the Selle and Meollion localities a thick Jurassic succession is preserved. At Fournel, however, basement is directly overlain by Eocene strata and towards the south, at Dourmillouse, the same situation recurs except that a gentle ESE tilt of the basement–cover unconformity introduces Triassic strata. No Jurassic rocks outcrop at Dourmillouse.

There is, therefore, a pre–Nummulitic basement high towards the west of the Fournel locality. Both Jurassic extensional faulting and pre–Eocene thrusting can be proposed to account for this (Figs.4.24, 4.25 & 4.26) :

1) a major basement fault downthrowing to the NW. It can be argued that the Fournel and Dourmillouse outcrops represent the uplifted footwall shoulder of a tilted basement fault block. In this model, the lower Jurassic sequence was initially condensed or absent on the uplifting footwall shoulder, which was gradually eroded to expose the Triassic and eventually the basement strata. Hence towards the east of Dourmillouse the Triassic is still preserved, but towards the west absent. Meanwhile at Selle, in the hangingingwall, a thick sequence of Liassic syn–rift sediments is preserved.

2) a major W–vergent basement thrust. According to this model the Mesozoic outcrop at Selle may lie in the footwall of a basement thrust. A pre–Nummulitic basement thrust truncates the Selle syncline. It can be postulated that this structure continues along strike and produced significant uplift which caused the erosion of the Mesozoic cover sequence from its hangingwall. The Selle syncline is thus considered as a footwall syncline to a major pre–Nummulitic basement thrust that is hidden beneath the drape of flysch sediments.

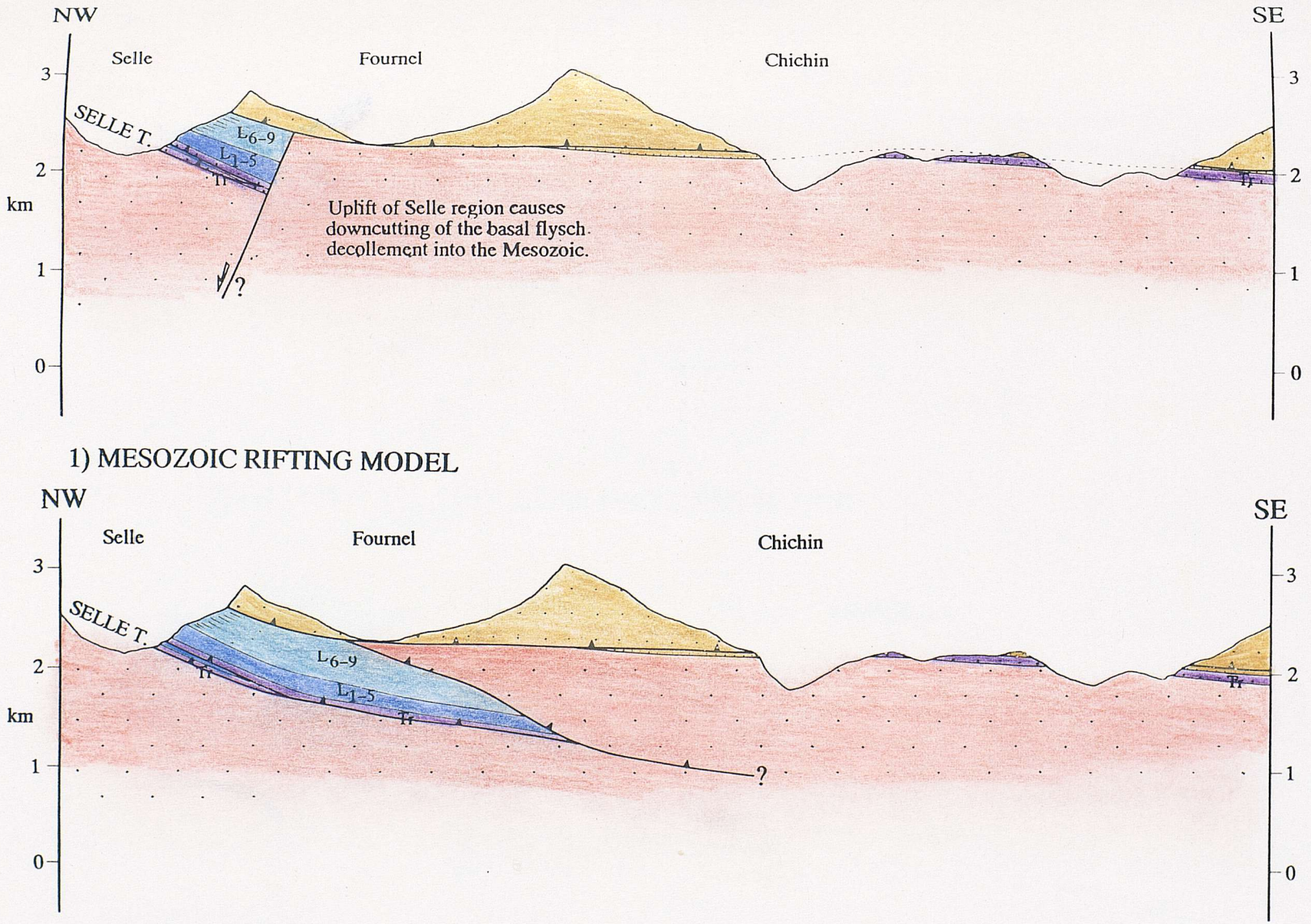
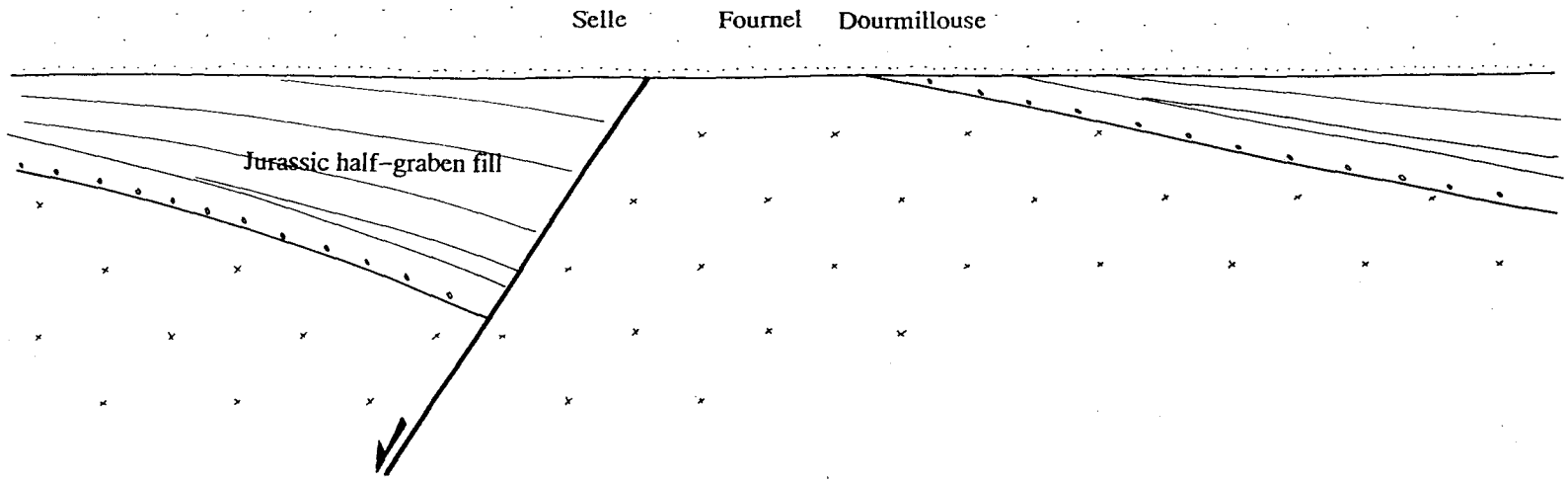


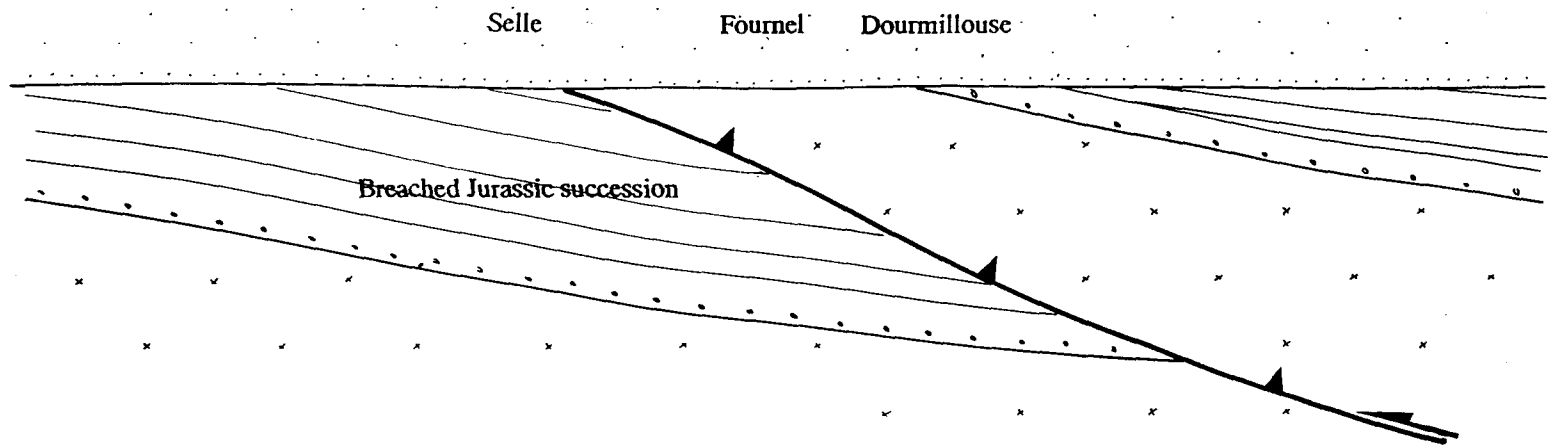
Fig.4.25 SECTION ACROSS THE SELLE, FOURNEL AND DOURMILLOUSE STRUCTURES,

showing the uplift of the Nummulitic unconformity at Selle and two models of the pre–Nummulitic structure

Fig.4.26 CARTOON ILLUSTRATING THE POSSIBLE PRE–NUMMULITIC STRUCTURE BETWEEN SELLE AND DOURMILLOUSE, and the absence of Mesozoic strata at Fournel.



1) UPLIFTED FOOTWALL SHOULDER OF TILTED BLOCK

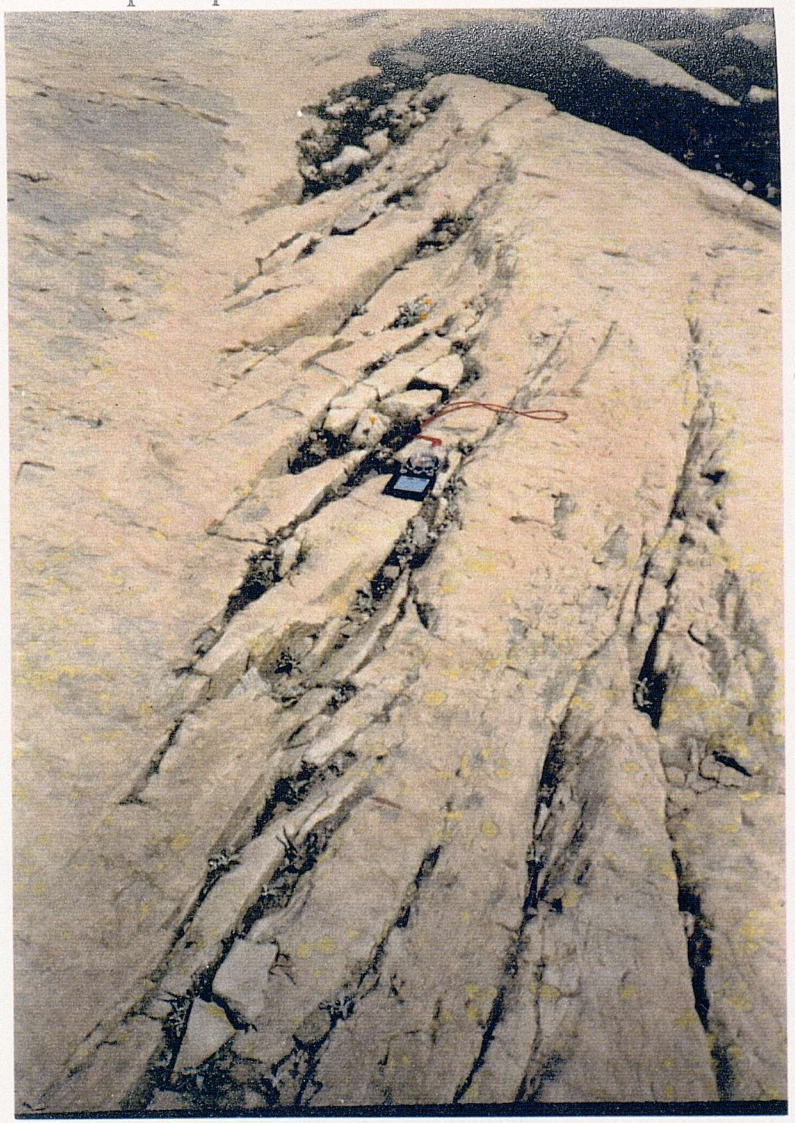


2) UPLIFTED HANGINGWALL OF BASEMENT THRUST

possesses a strong lineation to N235°E



principal stress direction N103°E



EXAMPLES OF TRANSPORT DIRECTION INDICATORS FROM DOURMILLOUSE,

showing a) a shear zone within Triassic spilite

b) large scale tension gashes in Eocene flysch

Satisfactory stratigraphic evidence to distinguish between these two models is unavailable. The latter model is favoured by the author due to the presence pre–Nummulitic basement folding and thrusting at Selle.

4.3. FIELD DATA FROM EASTERN PELVOUX.

The localities examined in eastern Pelvoux lie in the environs of Vallouise (Fig.4.27). Vallouise lies in the Gyronde valley which is reached by car from l'Argentiere–la Bessée. Directly north of Vallouise lies the Chambran valley which rises to the peak of l'Yret (2830m). West of Vallouise lies La Blanche (2954m). Between these peaks a thin ribbon of Eocene flysch is exposed, bounded by the Brianconnais zone to the east and the Hercynian basement of the Pelvoux massif in the west.

No Mesozoic strata are preserved in this locality. Rather the Eocene series lies unconformably upon the Hercynian basement suite. There was, therefore, significant pre–Eocene uplift that caused the erosion of the Dauphinois zone cover sequence. Evidence of post–Eocene deformation is abundant; the Eocene unconformity is strongly folded at La Blanche, the basal Eocene beds are mylonitised at l'Yret and late normal faulting has affected both the Nummulitic flysch series and the Brianconnais shelf sediments.

4.3.1. Chambran–l'Yret.

The Chambran outcrop is quite unlike the localities previously described. The proximity of the Frontal Pennine Thrust (FPT) caused the development of a much more intense shear deformation in post–Eocene times. As one moves northward from Vallouise to l'Yret summit the preserved thickness of the Nummulitic flysch series decreases markedly from 1000m+ to less than 20m (Fig.4.28). This is caused by lateral ramping of the FPT. Northwards from l'Yret the flysch sequence thickens again.

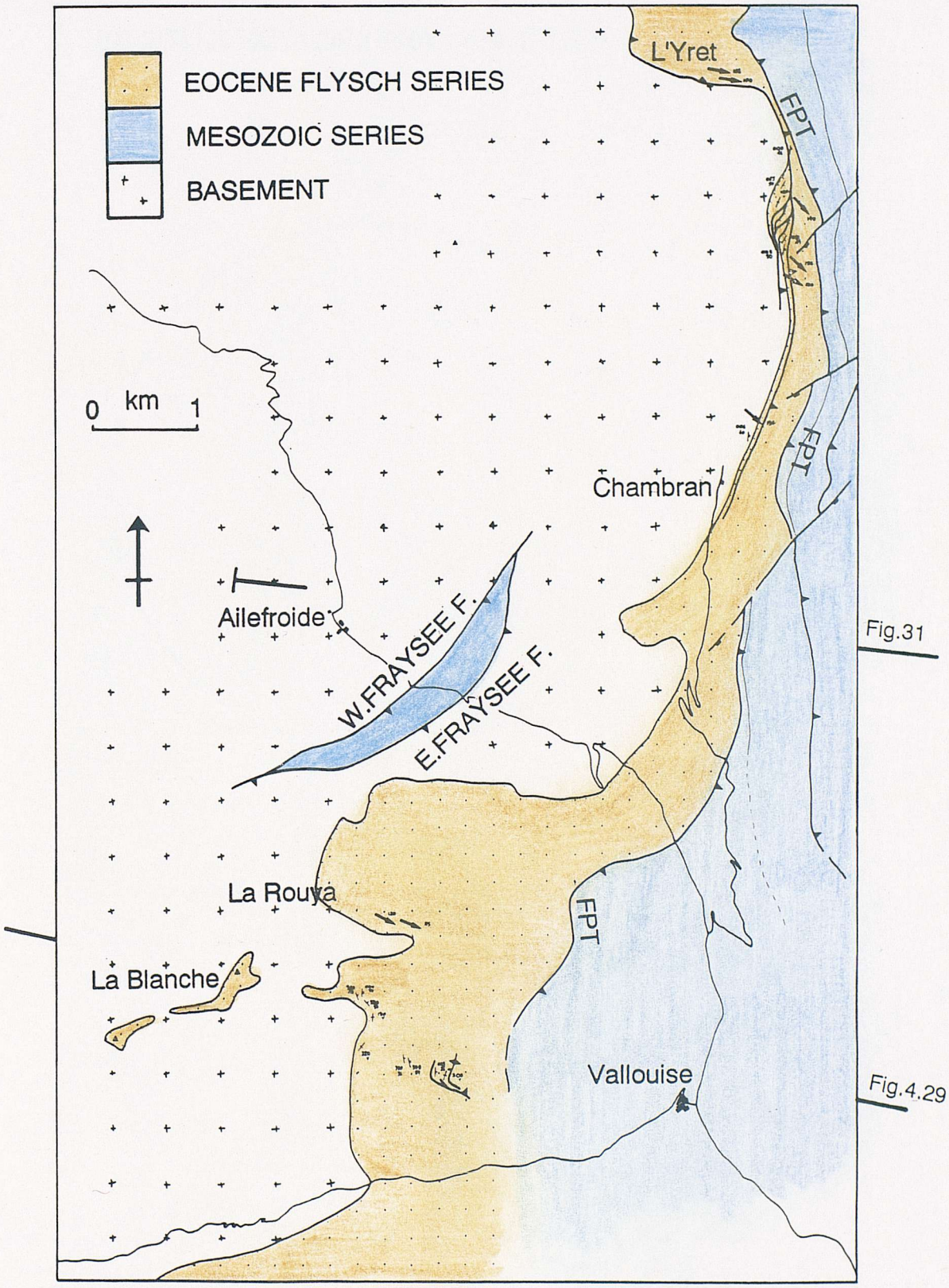
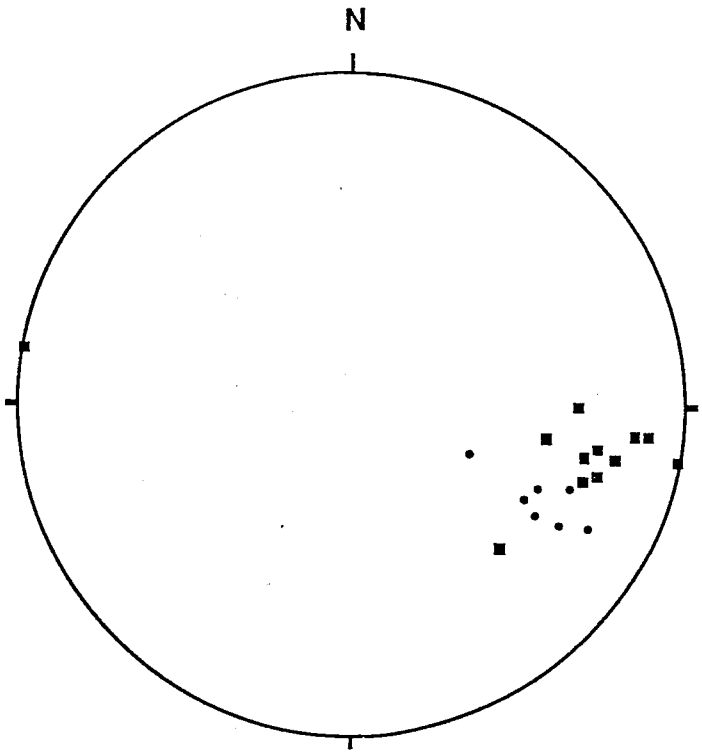


Fig.4.27 MAP OF THE LA BLANCHE, CHAMBRAN–L'YRET AND AILEFROIDE LOCALITIES, EASTERN PELVOUX.

Mineral Lineation data from Eastern Pelvoux



Key

- ■ 11 Data from Chambran/L'Yret
- • 7 Data from La Blanche

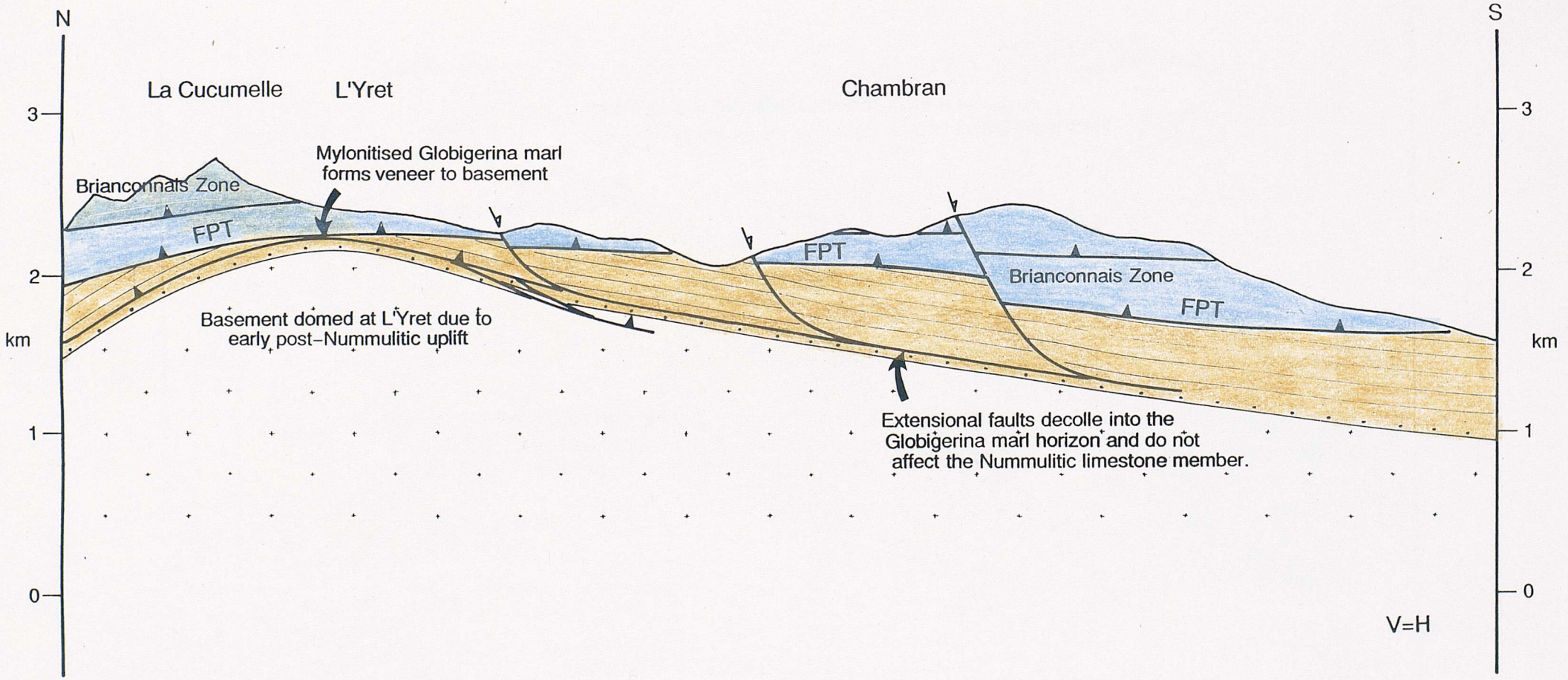


Fig.4.28 STRIKE-SECTION THROUGH THE FPT ZONE AT CHAMBRAN,

showing the uplift of basement at L'Yret and the collapse of the Brianconnais Zone into the Eocene flysch.

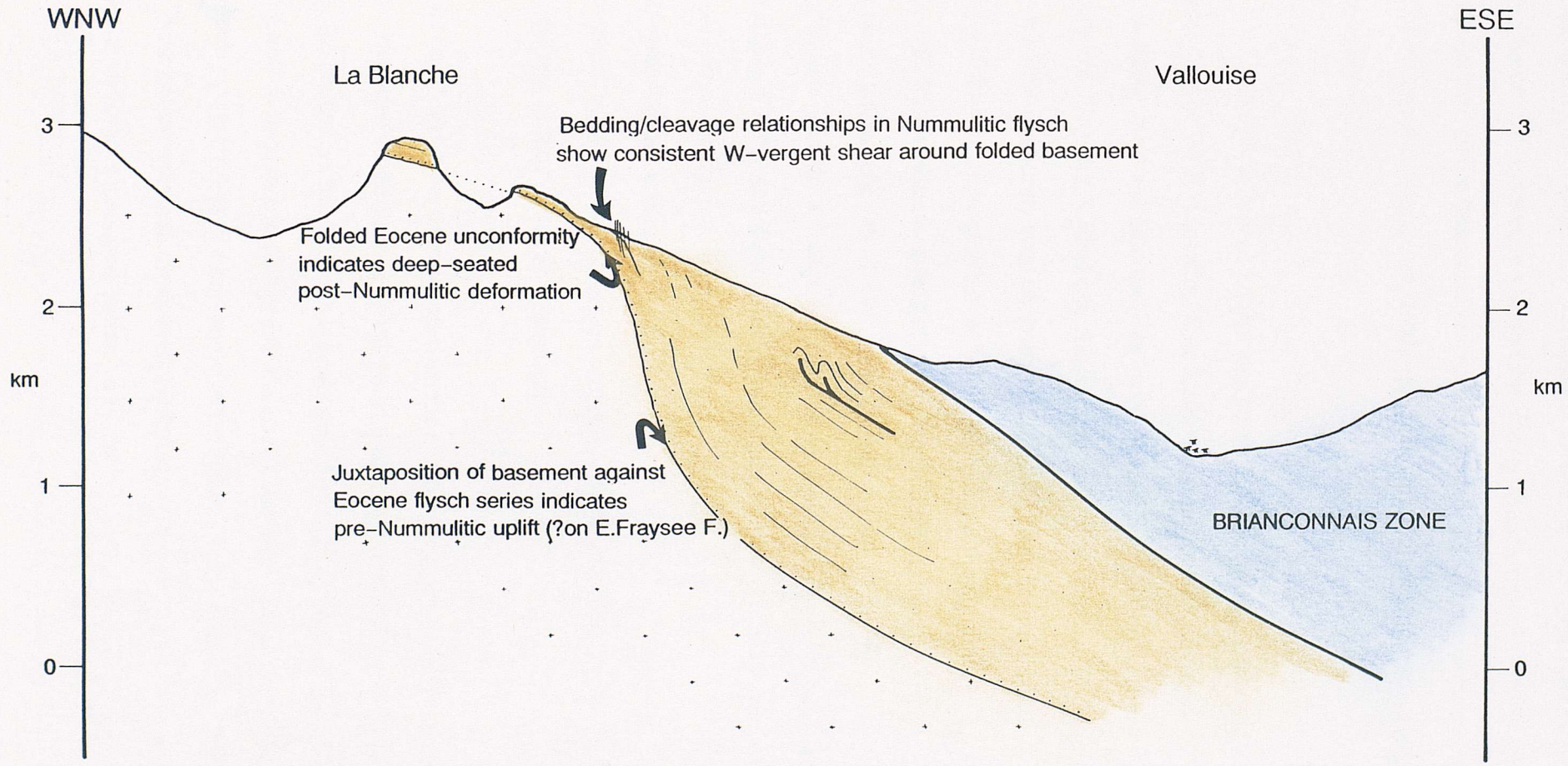


Fig.4.29 SECTION THROUGH LA BLANCHE,

showing post–Nummulitic folding of the Eocene unconformity and shearing in the footwall to the FPT.

For location see Figure 4.27

East of Chambran village the Nummulitic limestone lies upon basement gneisses and although the limestone is strongly sheared the stratigraphic unconformity is preserved. Further north the topography of the basement surface is revealed; its slightly undulose nature being reflected by the Nummulitic limestone bedding planes which are scalloped, yielding highly variable bedding orientations.

Cleavage foliation within the basal Eocene units strikes approximately N–S (N006°E) and dips at 60°E. Lineation data from the same locality suggests NW–vergent shear (N134°E). Bedding–cleavage intersections trend NNE–SSW (N195°–210°E) however and dip gently southward. Between Chambran and the summit of l'Yret the Eocene unconformity is imbricated. Mineral lineations vary between N120°–140°E in these imbricates, but the strike of the cleavage foliation remains constant at N000°–008°E. At the summit of l'Yret the shearing is most intense. Mylonitic fabric is manifest by the basal marly unit of the flysch series, the strong mineral lineation indicating W–WNW shear (N100°–103°E). At this point the Nummulitic limestone is absent and the Globigerina marl is juxtaposed against the Hercynian basement. There has therefore been stratal omission – not stratal duplication as is normally observed in compressional belts – indicating either :

a) down to the east extension,

b) the development of west–vergent breaching thrusts or

c) variations in the Numulitic stratigraphy.

Late normal faults have been mapped on the eastern flank of the Chambran valley. These structures are restricted to the Eocene flysch series and do not affect the basal Nummulitic limestone unit or the Hercynian basement. It appears therefore that the faults decolle in the Globigerina marl unit and accommodate collapse of the overlying flysch series and Brianconnais thrust stack. Downthrow on the faults is consistently to the south. In this direction the preserved thickness of the flysch series increases away from the basement high of l'Yret. It seems that the load of the Brianconnais thrust sheets was supported by the basement high but caused collapse where the flysch basin is preserved.

4.3.2. La Blanche

On the slopes of La Blanche, towards Vallouise, the Eocene unconformity is well-exposed in the footwall of the FPT. At the Rif des Estettes the unconformity dips subvertically, but at the Crete des Barres and at the summit of La Blanche the unconformity flattens to 25-35 degrees (Fig.4.29). The region therefore experienced post-Eocene uplift and folding on deeper basement structures. However at surface one observes higher level shears and folds within the Eocene flysch sequence.

These folds are assymetric, verging towards the WNW (subvertical bedding striking N200°E) and associated with a cleavage that strikes N-S (N348°-000°E) and dips at 70-90°E. The bedding-cleavage relationship consistently indicates west-vergent shear that predated the bulging of the Eocene unconformity.

Timing of the overthrusting of the FPT with respect to the bulging of the Eocene unconformity - both of these events being post-Eocene - is uncertain.

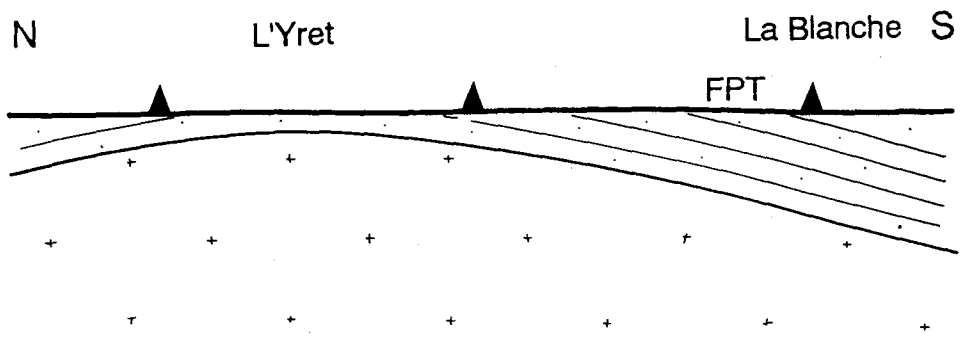
1) If the late FPT breaching system is assumed to be relatively planar then the pre-FPT basement topography could be constrained by the thickness of Eocene flysch preserved in the footwall of the FPT. This method suggests uplift of the basement at l'Yret by approximately 1000m relative to La Blanche.

2) However it can also be proposed that the absence of Eocene flysch at l'Yret is due to downcutting (Fig.4.30). The bedding cut-offs produced by both of these models (truncation of an uplifted dome and active downcutting of the FPT) are indistinguishable. However the present-day basement topography is strongly domed (Fig.4.28); the basement being uplifted by more than 1000m at l'Yret compared with Vallouise. The FPT on the other hand is more planar and would in fact maintain a constant level but for the presence of late extensional offsets. This

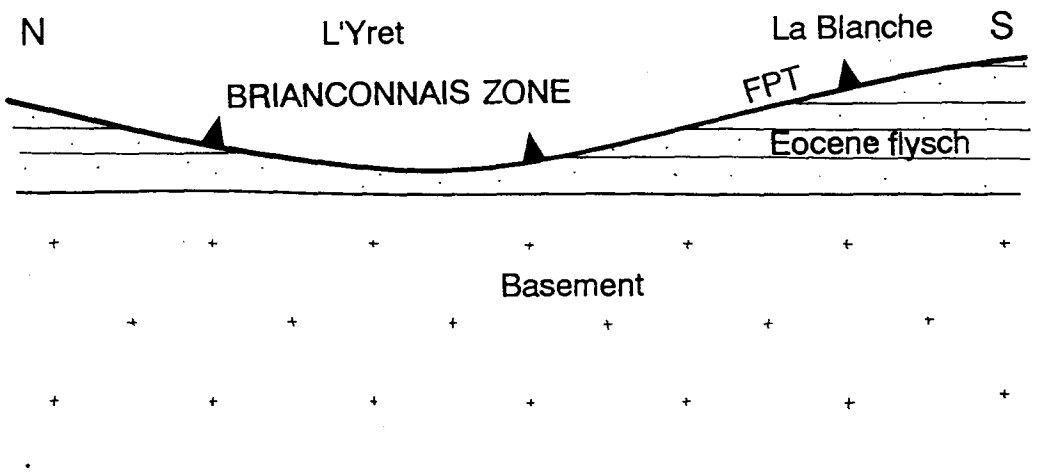


VIEW OF NUMMULITIC LIMESTONE UNCONFORMABLY OVERLYING BASEMENT NEAR THE SUMMIT OF LA BLANCHE

1) FPT TRUNCATES UPLIFTED BASEMENT DOME.



2) FPT CUTS DOWN LATERALLY INTO FLYSCH BASIN.



3) DIP-SECTION SHOWING FRONTAL RAMPING OF FPT TOWARDS THE WEST.

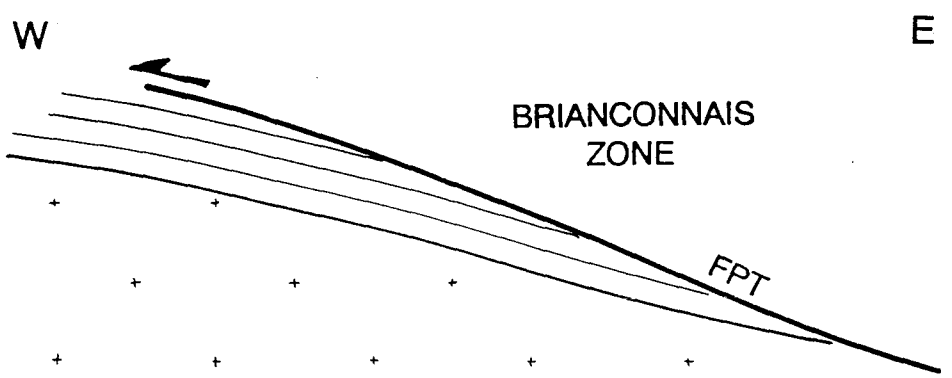


Fig.4.30 CARTOONS ILLUSTRATING THE RELATIONSHIP BETWEEN THE EOCENE FLYSCH SERIES AND THE FPT.

suggests that the basement was uplifted in early post–Eocene times prior to the propagation of the FPT, which truncates the basement dome.

4.3.3. Ailefroide.

Between Vallouise and Ailefroide Mesozoic sediments are exposed in a small outcrop that straddles the Gyronde valley. The intensely deformed sediments are sandwiched between basement rocks in both the east and west (Fig.4.27). Towards the east the basement is unconformably overlain by the Eocene trilogy.

On the northern slopes of La Rouya (2750m) the lower Jurassic succession is inverted in the footwall of a basement thrust. This is also clear from the structural relationships mapped to the north of the Gyronde River at Rif de Fraysee (Fig.4.31). Here, the bedding–cleavage relationship in footwall of the basement overthrust (the E.Fraysee Fault) indicates overturned bedding and this is supported by bedding cut–off data at the thrust contact. Bedding–cleavage lineations trend N160°–204°E and dip at 20–37°S. The bedding data is tightly gathered, striking N–NNE (N000°–032°E) and dipping at 38–55°E. This indicates that the folding, manifest by cleavage–bedding relationships in the absence of observable fold hinges, was tight to isoclinal.

Towards the west the stratigraphy is the right–way–up. Triassic beds are omitted however. The basement–cover contact is intensely sheared and basement is juxtaposed successively against lower calcareous Lias and td then upper shaley Lias units as one follows the contact south. The original nature of this western basement–cover contact is uncertain. The NE–SW strike of the contact and the omission of Triassic and then lower calcareous Lias sediments is consistent with an original extensional fault geometry (the W.Fraysee Fault). It is possible, however, that the contact is simply a much sheared stratigraphic unconformity between basement and Trias rocks.

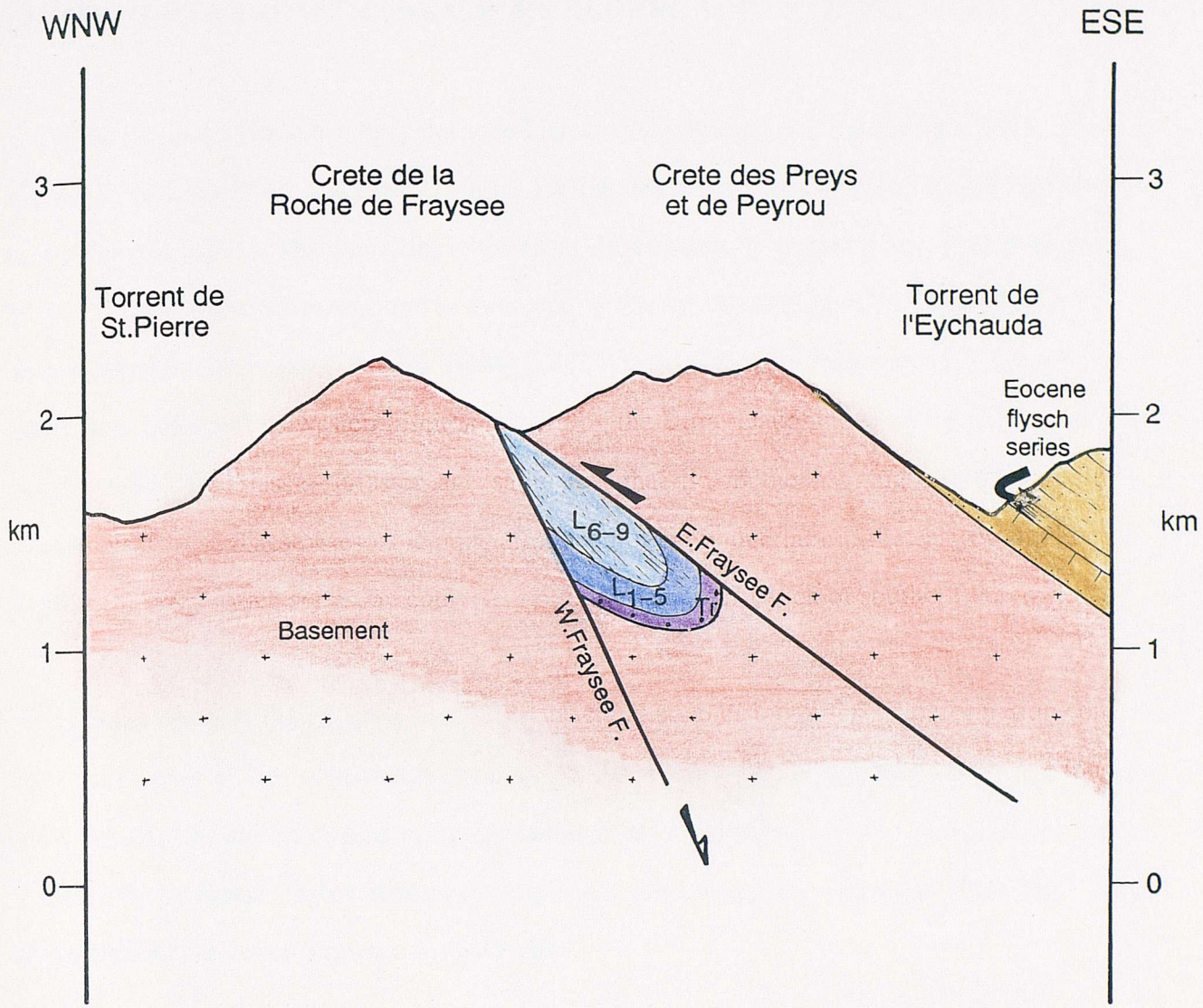


Fig.4.31 SECTION THROUGH THE AILEFROIDE LOCALITY, EASTERN PELVOUX, showing the pinched-in syncline of Mesozoic strata between the W.&E.Fraysee Faults.

4.4. FIELD DATA FROM THE ARVINCHE REGION, NORTH OF PELVOUX.

The Arvinche chain has been described by several workers, notably Barbier (1956, 1963), Barbier & Barfety (1972) and Beach (1981a, 1981b). More recently the area has also been studied by Trudgill (1991). The chain lies in the Ultra-Dauphinois Zone and is bounded in the south by the Pelvoux basement massif and in the north by the Arc River (Fig.4.32). Its eastern boundary is marked by the Frontal Pennine Thrust (FPT) which carries the Brianconnais Zone over the Dauphinois Zone and its western limit is marked by the Ultra-Dauphinois Thrust. The Arvinche chain consists of Triassic-Jurassic age sediments that lie in the hangingwall of the Ultra-Dauphinois Thrust. These Mesozoic units are unconformably overlain in the east by the Eocene Aiguilles d'Arves flysch series, an equivalent of the Champsaur flysch of southern Pelvoux.

To the north of the Pelvoux massif the structural style of the observed deformation contrasts markedly with that observed in the massif itself. This is due to the structural level exposed at the surface; in Pelvoux Alpine uplift has resulted in the unroofing of the Hercynian basement whilst in the Arvinche region basement suite rocks remain unexposed and only the high level deformation of the cover sequence is observed.

Differing views of the deformation in the Arvinche chain have been postulated by Barbier (1956, 1963) and Beach (1981).

1) Barbier (1956) regards the base of the Eocene series as a stratigraphic unconformity and therefore argues that structures that rise to meet the unconformity are truncated, and are therefore pre-Eocene. Major pre-Eocene uplift is fundamental to Barbier's (1956) concept of the Arvinche chain. He then envisages a relatively minor phase of post-Eocene deformation.

2) Beach (1981), on the other hand, suggests that the Eocene unconformity has been strongly sheared and that the main deformation event was, in fact, post-Eocene.

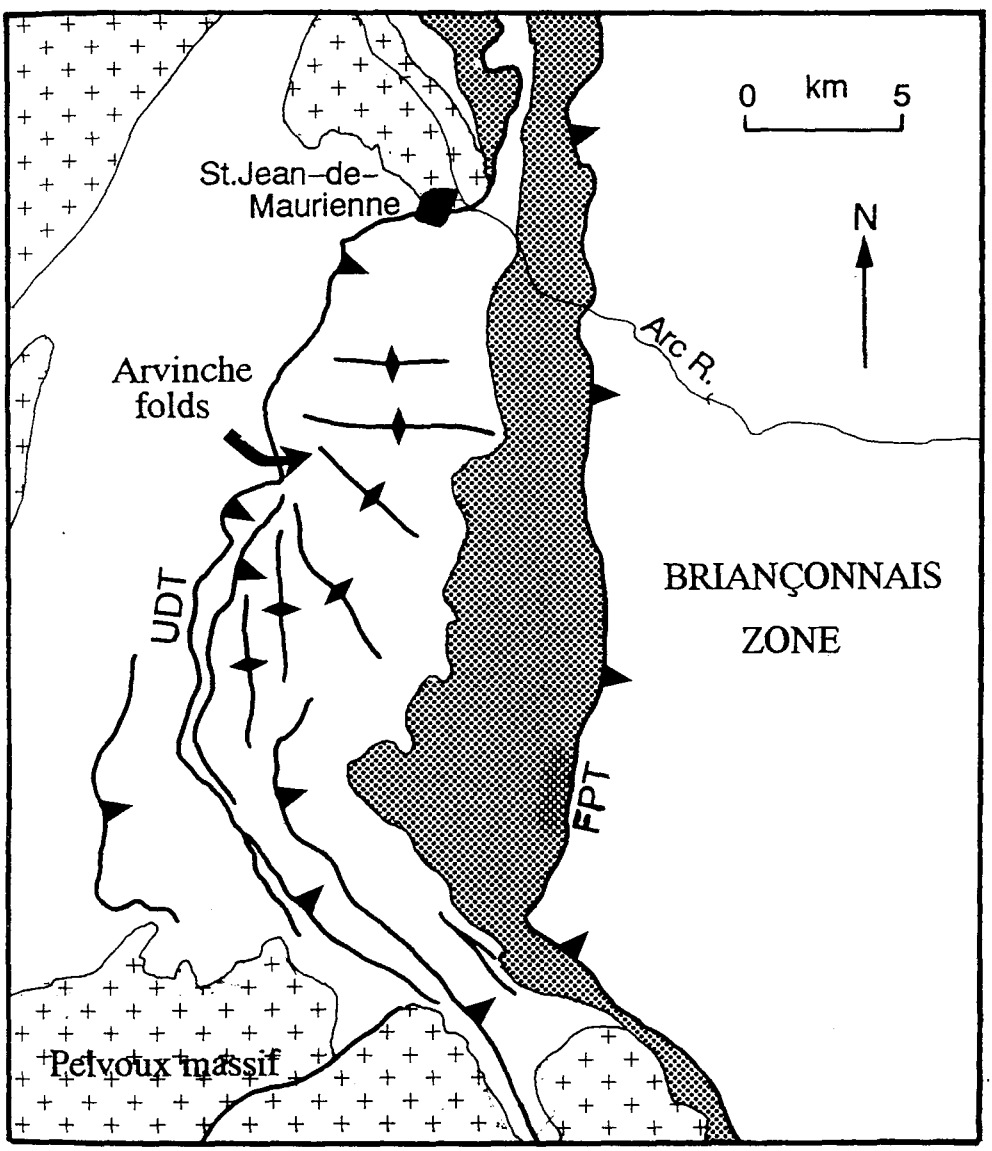


Fig.4.32 MAP OF THE ARVINCHE REGION, NORTH OF PELVOUX.

UDT, Ultra–Dauphinois Thrust; FPT, Frontal Pennine Thrust; dotted area, Eocene flysch; blank area, Mesozoic sediments; crossed area, external basement massifs.

Fieldwork carried out by the author indicates that both pre- and post-Eocene deformation has affected the Arvinche region. The pre-Eocene phase is characterised by N-vergent folding in the Mesozoic sequence that is truncated at the Eocene unconformity. The style of folding developed in the Triassic evaporite and lower calcareous Lias units is markedly different from that developed in the middle Jurassic carbonates. This disharmony is facilitated by the 750m+ thick upper Lias-lower Dogger shaley sequence. The post-Eocene phase is represented by W-vergent thrusts and associated folds. The Ultradauphinois Thrust breaches the Eocene flysch to the north of the Arc River. Other W-vergent thrusts rise to meet the Eocene unconformity near the Aiguilles d'Arves - these thrusts being the subject of debate between Barbier and Beach. These imbricates occur in the south where Triassic evaporites are absent, but towards the north the existence of E-W trending, gypsum-cored pre-Eocene folds prevents the development of W-vergent thrusts in the hangingwall of the Triassic-hosted Ultradauphinois Thrust.

The stratigraphic column exposed in the Arvinche chain is more complete than, and distinct from, that of Pelvoux (Fig.4.33). The oldest exposed strata are of Triassic age. These are well-developed gypsum/anhydrite deposits that form a major slip horizon exploited by the Ultradauphinois Thrust. This salt basin is replaced towards the south by dolomites. The lower Lias facies in Pelvoux and the Arvinche are identical, but the sequence thickens towards the north. Subsidence was also greater in the Arvinche throughout upper Lias and lower Dogger times, when a thick shale series was deposited. More calcareous conditions then returned. Pre-Eocene uplift eroded any upper Jurassic and Cretaceous strata that were deposited. The Eocene sediments lie unconformably upon the Mesozoic series. The Eocene Aiguilles d'Arves flysch series is equivalent to the "trilogie Nummulitique" of Champsaur. In Arvinche the Nuummulitic limestone unit is not well-developed, the series consisting of shale overlain by flysch. The basal conglomerate at the Aiguilles d'Arves becomes particularly massive, exceeding 400m thickness.

The Arvinche region can be considered in two parts.

1) Firstly, **the Triassic-Jurassic exposures in the hangingwall of the Ultradauphinois**

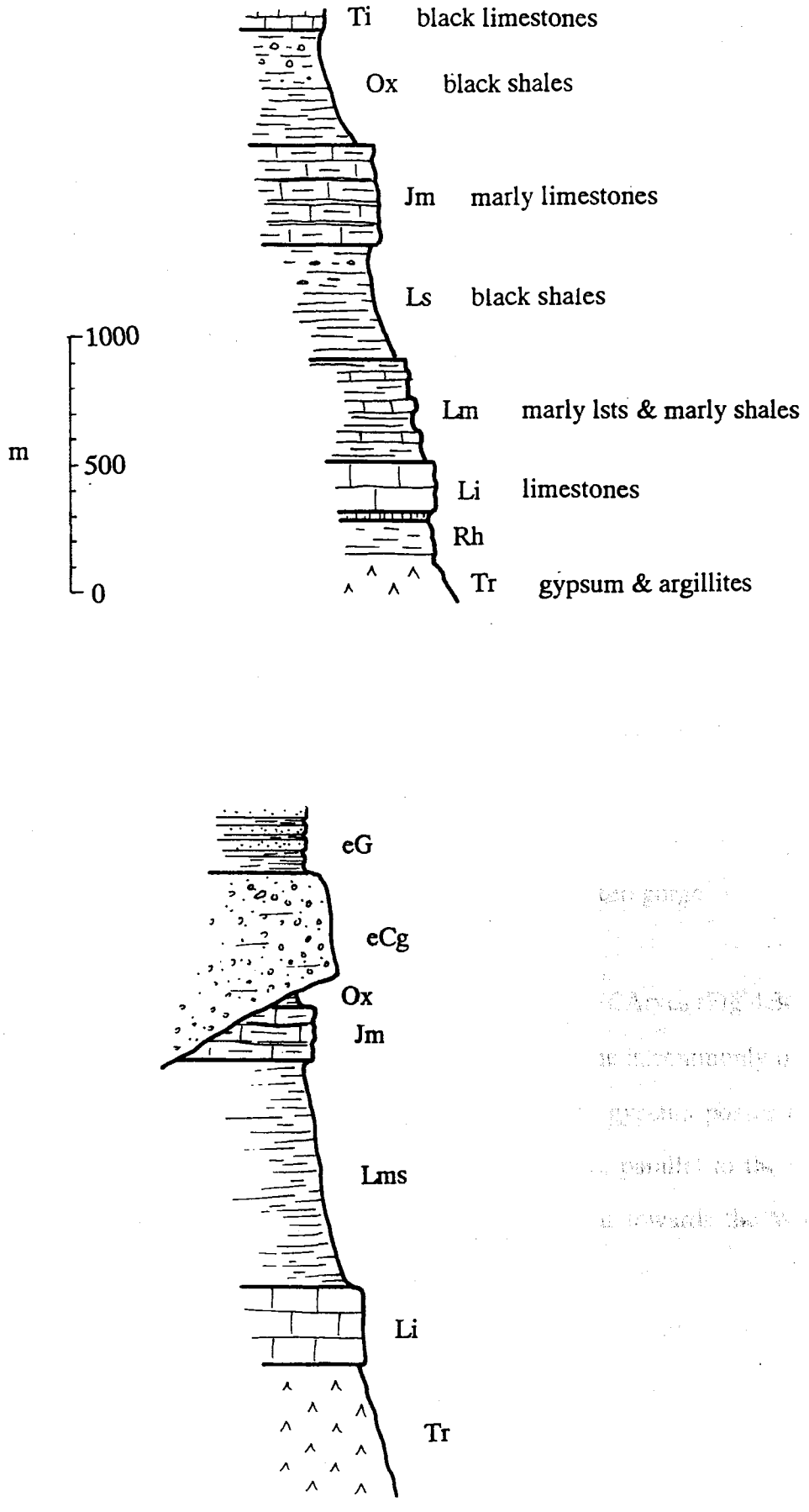


Fig.4.33 STRATIGRAPHIC COLUMNS FROM THE ULTRA-DAUPHINOIS THRUST SHEET (from BARBIER (1956))

Thrust will be discussed, including outcrops from the Arvan valley and the western flank of the Aiguilles d'Arves ridge.

2) Then **the structures exposed in the Eocene flysch** to the east of the Aiguilles d'Arves ridge are described.

4.4.1. The Arvan Valley.

The Arvan River flows northward to join the Arc valley at St.Jean-de-Maurienne, incising deeply into the Mesozoic strata. The principal exposures are reached from the road (D.926) between St.Jean and the village of Entraigues. Access to the western flank of the Aiguilles d'Arves ridge is via the road (D.80) to Albiez-le-Vieux from just northeast of Entraigues.

The topography of this region very clearly reflects the underlying geology. The Aiguilles d'Arves ridge is capped by resistant Eocene conglomerates and sandstones. These unconformably overlie the thick upper Lias-lower Dogger shaley sequence that has been eroded to form a gently undulating elevated pasture. In the west, the Arvan River has cut rapidly through the folded Triassic gypsum and lower Lias limestone formations, producing a deep gorge.

The Ultradauphinois Thrust is exposed to the north of St.Jean-d'Arves (Fig.4.34) where it carries Triassic gypsum over upper Lias black shales. The thrust plane is commonly obscured by scree but can be observed in stream sections. Close to the fault the gypsum possesses a strong shear foliation striking approximately N010°E and dipping at 32°E, parallel to the fault plane itself. Mineral lineations in the subjacent Lias shales indicate shear towards the WNW, their azimuths varying between N096°-123°E. The Liassic strata have been folded and associated cleavage foliations strike N-S (ranging from N342°-022°E, the mean strike being N008°E) and fan strongly through moderate to subvertical dips. The bedding cut-off data indicates that the Ultradauphinois Thrust cut up-section towards the WNW-NW, the Triassic gypsum successively overriding Domerian-Aalenian shales, Dogger shaley limestones and eventually the Eocene flysch. **The combined data indicates WNW-vergent overthrusting on the Ultradauphinois**

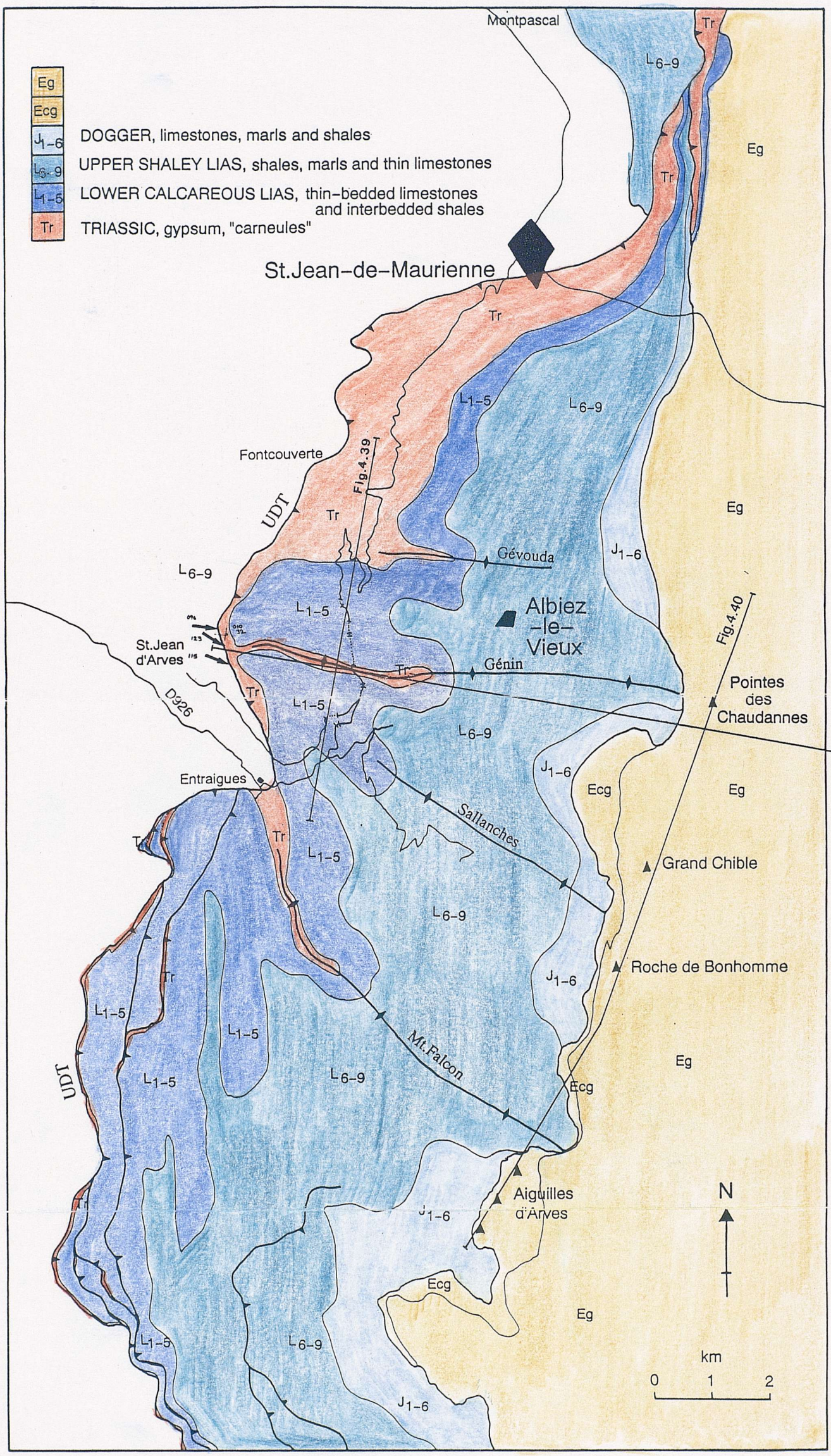
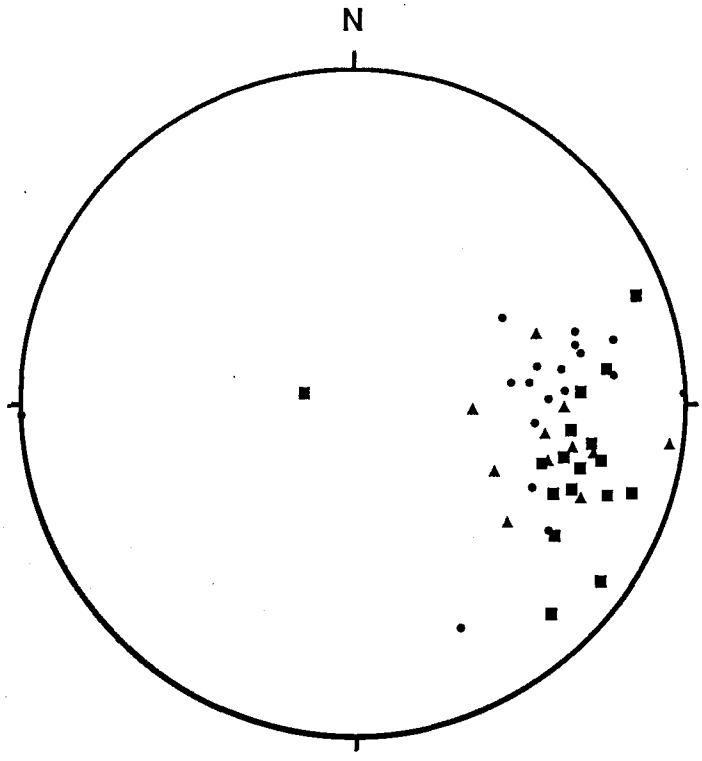


Fig.4.34 MAP OF THE ARVINCHE FOLD SYSTEM.

UDT, Ulta-Dauphinois Thrust
Showing location of Figures 4.37 & 4.39

Mineral Lineation Data from North of Pelvoux



Key

▲ 17 Data from La Cote Rouge

• 17 Data from the UDT

■ 11 Data from Aiguilles d'Arves

Thrust.

Towards the south, the Triassic changes from the evaporitic gypsum facies to the dolomitic "cargneules" facies. This produces a marked change in structural style, since the increased basal friction associated with the cargneules facies generates a number of splays from the Ultradauphinois Thrust. In fact, no single Ultradauphinois Thrust plane can be mapped south of Entraigues. Rather, numerous thrust repetitions and folds of lower Lias strata outcrop. This is the source of the problem of how to trace the Ultradauphinois Thrust into the Pelvoux massif (B.R.G.M. 1:50,000 La Grave sheet).

Northwards, as already mentioned, the Ultradauphinois Thrust ramps from a footwall flat in the upper Lias shales to a flat within the Eocene flysch sequence. Small exposures of Eocene rocks are exposed in the footwall of the Ultradauphinois Thrust to the west of Fontcouverte, but good exposures are only found to the north of the Arc River. At Montpascal, the Triassic gypsum is thrusted over Liassic shales (Fig.4.35). The bedding-cleavage relationship in the footwall strata indicates that bedding is folded to form a tight footwall syncline. The cleavage dips shallowly eastward whilst the bedding dips more steeply at 40-45°E and strikes NNW-SSE (N330°-011°E). The mineral lineations within these footwall sediments are aligned approximately W-E (varying between N076°-092°E). This data suggests WSW-vergent thrusting on the Ultradauphinois Thrust at this locality.

Hence the Ultradauphinois Thrust movement direction appears to swing from WNW to WSW as one moves north along its outcrop. This convergence of the transport directions could explain the fanning outcrop of the Arvinche fold train (Fig.4.36). This model envisages the Ultradauphinois Thrust sheet, riding on a weak gypsiferous basal layer, and accommodating non-colinear transport directions by internal strain - reorienting the pre-existing Arvinche folds.

The Arvinche folds of Barbier (1956) are marvellously exposed in the gorge of the Arvan river, north of Entraigues (Figs.4.37 & 4.39). The folds trend E-W, verge towards both the north

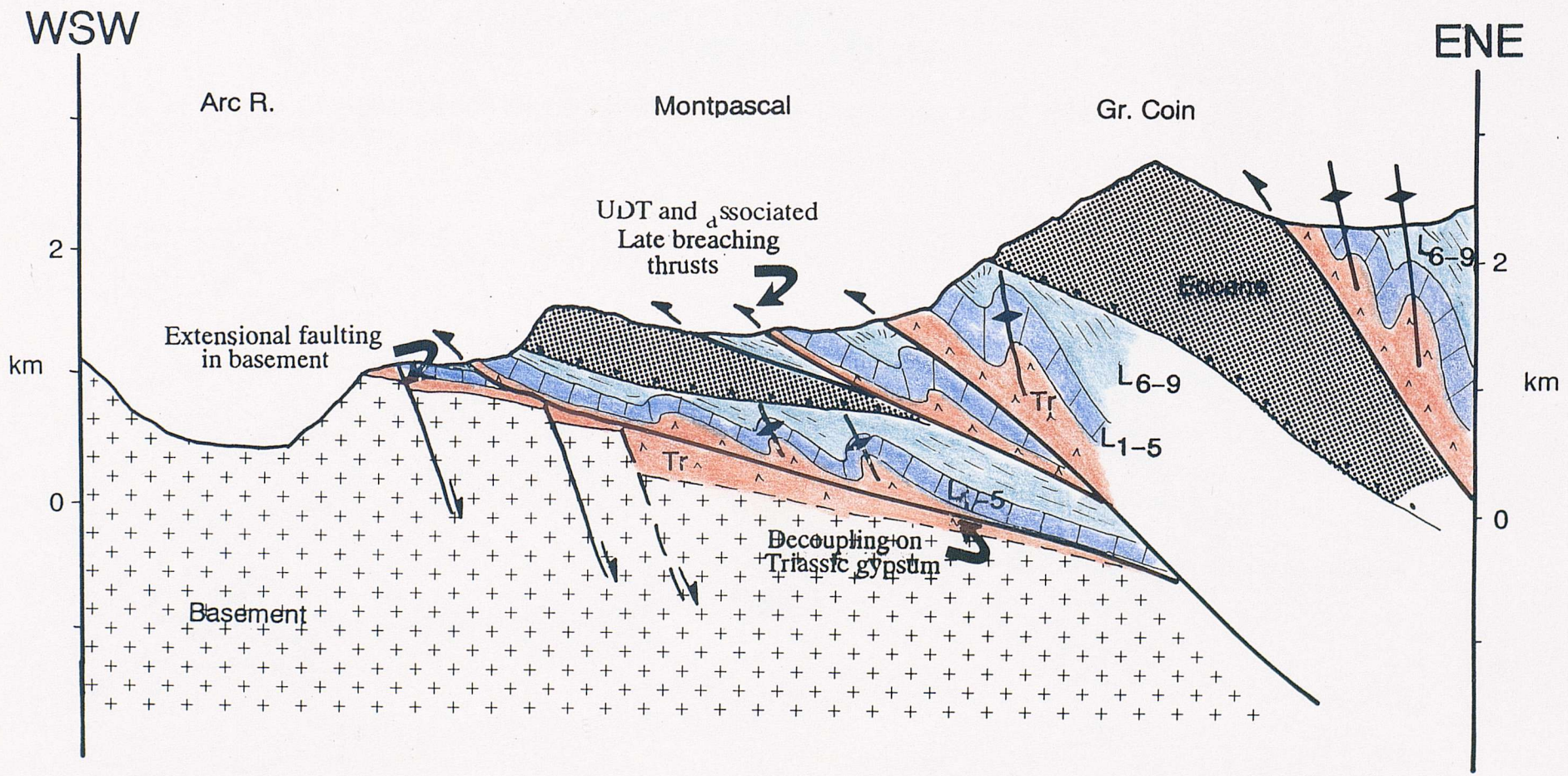
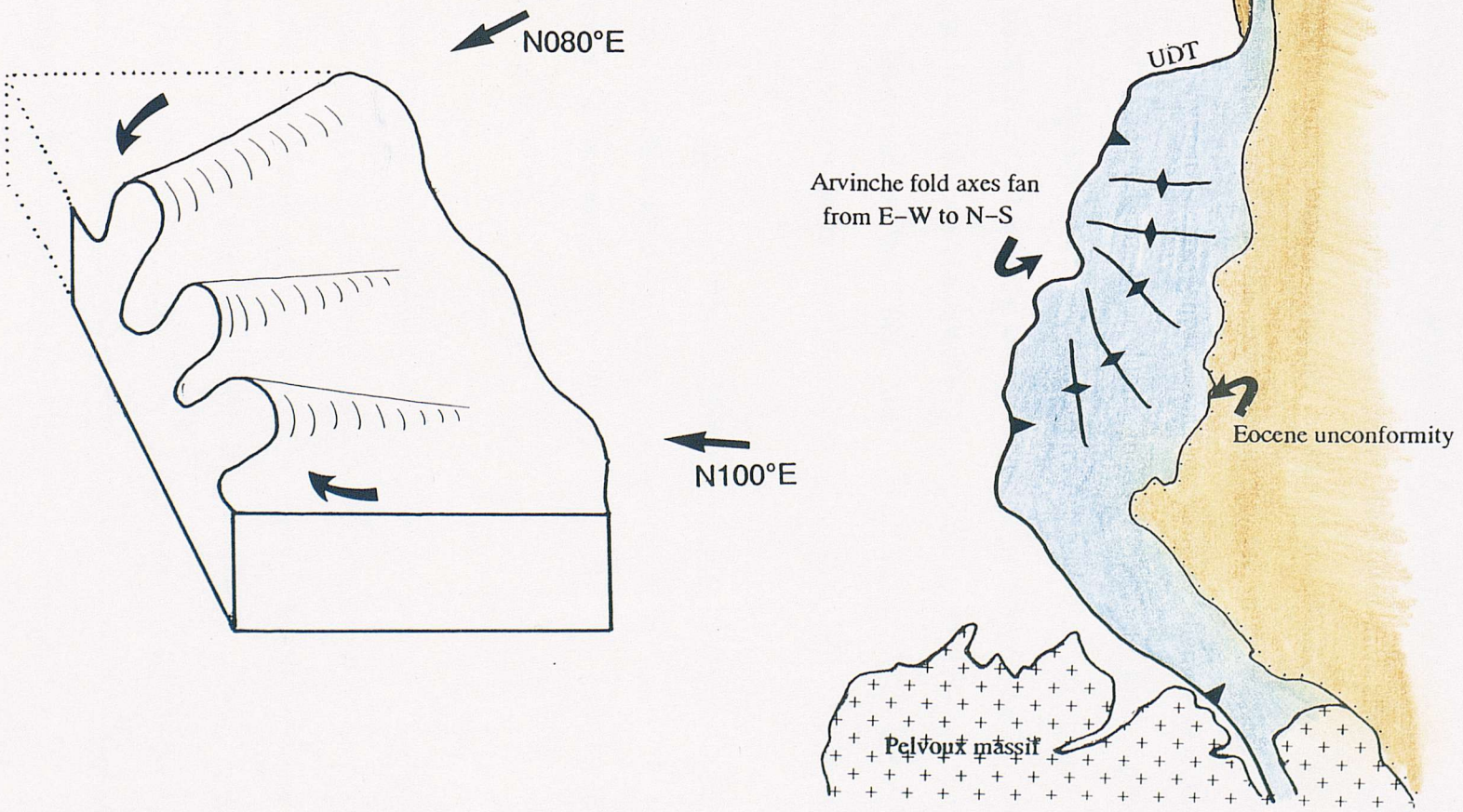


Fig.4.35 SECTION ACROSS THE ULTRA-DAUPHINOIS ZONE,

showing early folding truncated by Eocene flysch and late breach thrusting associated with the Ultra-Dauphinois Thrust.

Fig.4.36 CARTOON ILLUSTRATING A POSSIBLE ORIGIN OF THE FANNING OF THE ARVINCHE FOLDS,

non-coaxial transport directions within the Ultra-Dauphinois Thrust sheet result in along strike compression?



and south and plunge at approximately 30–35°E, a dip that correlates with the dip of the Ultra-dauphinois thrust plane. Towards the east the plunge of the folds may decrease.

The fold style varies with lithology. At low structural levels, broad flat synclines are separated by dramatic isoclinal anticlines that resemble diapirs. The folds are defined by the lower calcareous Lias formation which forms a coherent, resistant band 300–350m thick. The anticlines are cored Triassic gypsum, and it is this unit that dictates the structural style of the region. As a rheologically weak layer it is easily sheared and acts as a decollement, producing the Ultradauphinois Thrust, and is capable of flow, producing the isoclinal folds of the Arvinche chain. At a higher structural level the middle Jurassic limestones are folded into a series of N-vergent anticline–syncline pairs. Hence although the dramatic folds of the Triassic–lower Liassic strata are bivergent, the higher level folds are consistently N-vergent. The disharmony between the two structural levels is facilitated by the 750m+ thick upper Liassic shale succession (Fig.4.38). This incoherent band has been deformed by both pure and simple shear processes.

These structures are truncated against the Eocene unconformity and are therefore pre-Eocene in age. The oldest strata involved in the folding is of upper Jurassic age, but Cretaceous strata may have been deposited and later eroded. They may be equivalent to the E–W trending, N-vergent pre–Senonian folds of Devoluy.

Four major anticlines are mapped, that from north to south have been named Gevoudaz, Genin, Sallanches and Mt. Falcon (Figs.4.34, 4.39 & 4.40). The Gevoudaz anticline trends E–W, plunges east and verges north. The Triassic gypsum in its core has risen over 300m above its base level. The most dramatically exposed anticline is at Genin, the Triassic core of which has been strongly weathered to form a scar on the flank of Mt. Charvin, Combe Genin. The lower calcareous Lias and Triassic gypsum units have been uplifted by approximately 600m above their base levels in the hangingwall of the Ultradauphinois Thrust. The limestone beds dip vertically on the northern limb of the fold, whilst on the southern limb bedding dips at 50–60°S, defining a N-vergent anticline. From the D.926 road the shape of the Triassic fold cores can be

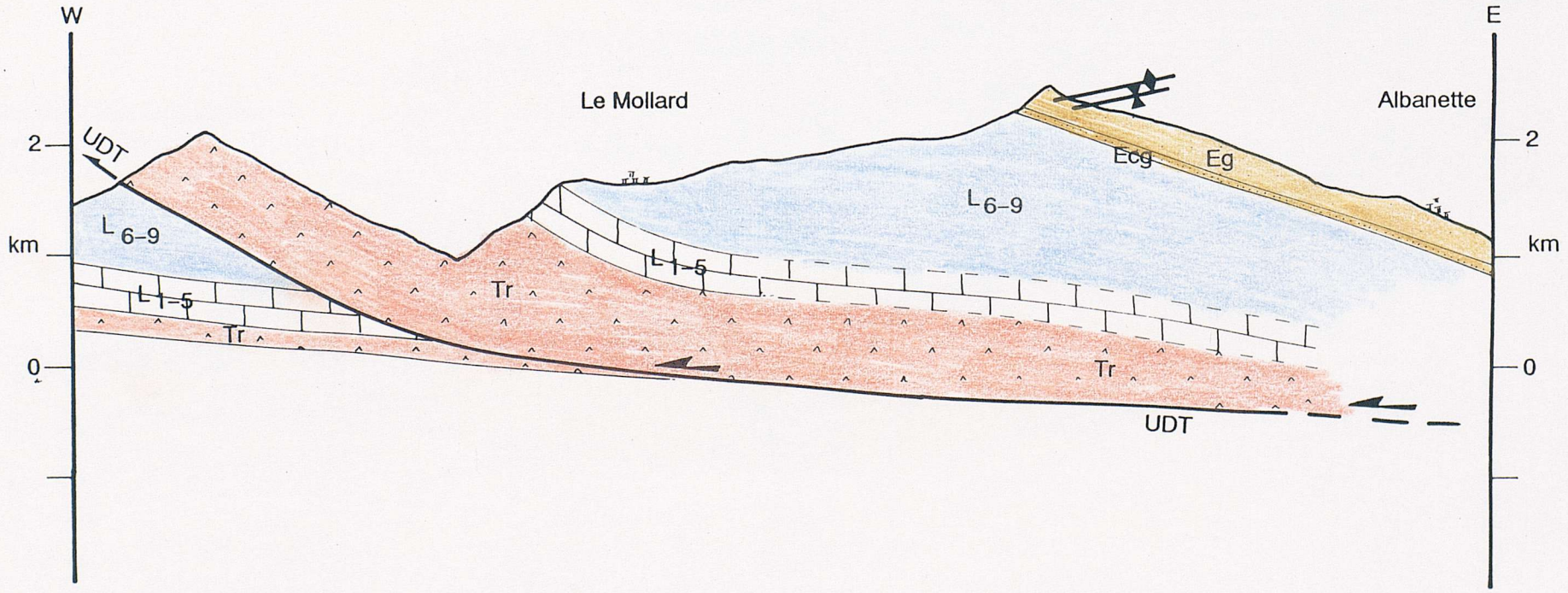


Fig.4.37 DIP SECTION THROUGH THE ULTRA-DAUPHINOIS THRUST
showing tilting of the Eocene unconformity towards the east and ramping of the UDT.

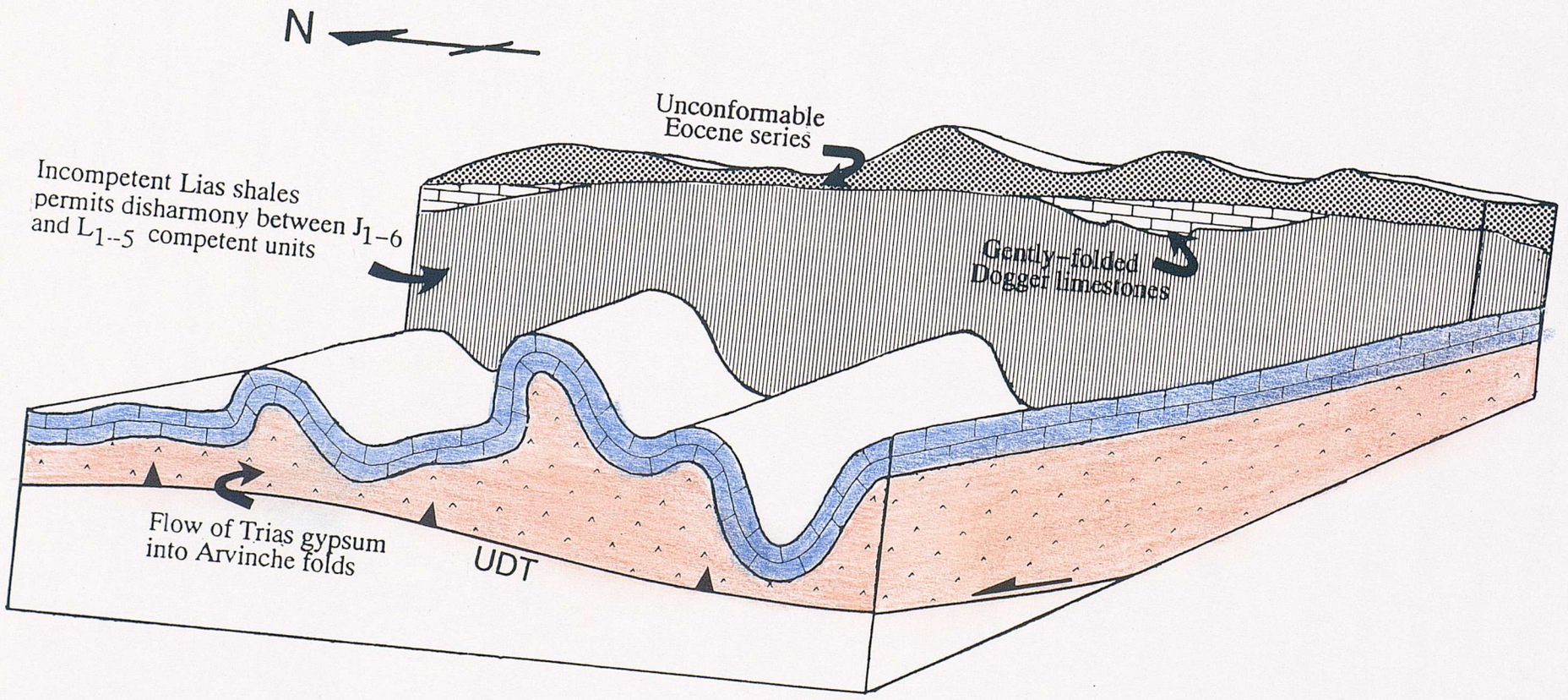


Fig.4.38 CUT-AWAY BLOCK DIAGRAM ILLUSTRATING THE STRUCTURE OF THE ARVINCHE REGION.

clearly observed on the eastern side of the Arvan gorge. The bedding and cleavage data that defines these folds is obtained from the the lower Lias limestones. Since the anticline cores on the road section have been strongly weathered, the plunge of the folds is obtained from the synclinal axes. The syncline between the Gevoudaz and Genin anticlines plunges at 18–25°E, although dips of up to 46°E are recorded. South of Genin, the syncline plunges at 30°E.

The Sallanche anticline verges south and trends ESE–WNW. The gypsum in its core fails to reach the heights of the Genin or Gevoudaz structures. However the syncline that separates the Sallanches and Mt. Falcon anticlines is particularly deep, suggesting extreme thinning of the gypsum in its core due to flowage. The upper Lias shales preserved in the core of this syncline are strongly cleaved, the cleavage striking NW–SE (N325°–335°E). This reflects the NW–SE trend of the Mt. Falcon anticline which is also south vergent.

The Arvinche fold axes fan from E–W to NW–SE orientations on the map. Various explanations can be proposed for this fan arrangement.

1) Since the Arvinche folds resulted from N–S directed compression it has been suggested that shortening might have increased towards the west.

2) Slightly convergent transport directions during the post–Eocene deformation phase may have reoriented the fold axes.

3) The fan–shape may also have occurred due to the loss of the gypsum facies towards the south. During post–Eocene compression the gypsum acted as a decollement, accommodating almost all displacement. However, towards the south this displacement is accommodated on a series of imbricates. Towards the south, strain is therefore taken up within the Ultradauphinois Zone rather than on the Ultradauphinois Thrust itself, and this strain has reoriented the fold axes. This effect increases towards the south and is particularly marked at Mt. Falcon.

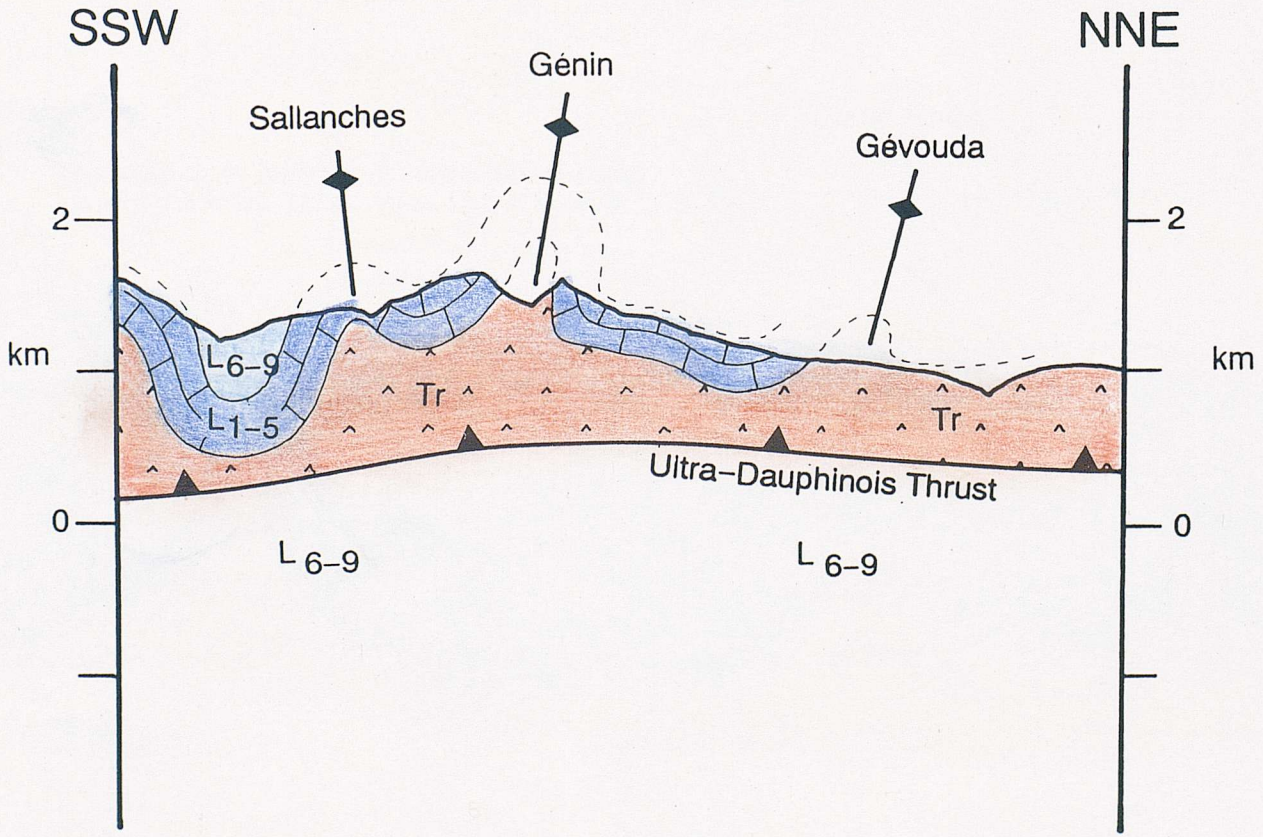


Fig.4.39 SECTION THROUGH THE ARVAN GORGE,
showing gypsum-cored anticlines in the hangingwall of the Ultra-Dauphinois Thrust. Transport direction of UDT sheet is into the page

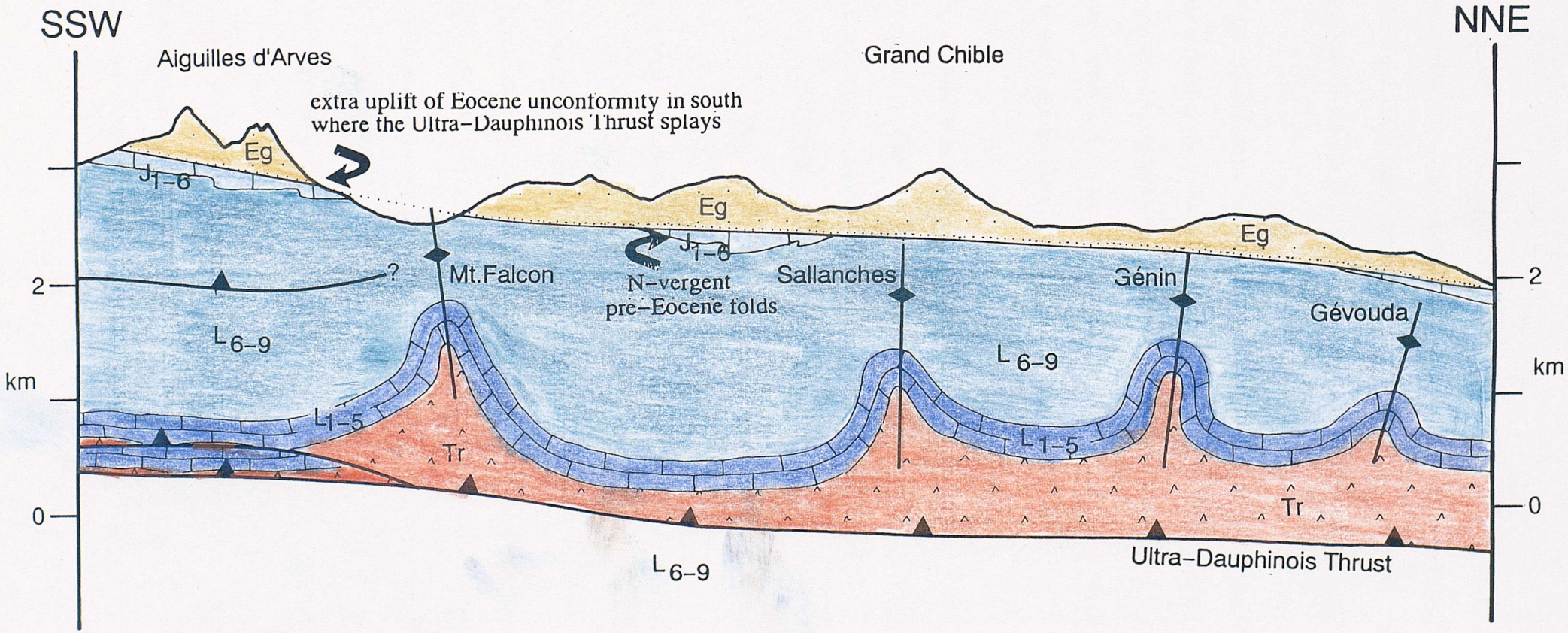


Fig.4.40 SECTION ALONG THE AIG. D'ARVES-GRAND CHIBLE RIDGE,
showing the Arvinche folds in the hangingwall of the Ultra-Dauphinois Thrust truncated by the Eocene unconformity. Transport direction of the UDT sheet is into the page

The folding observed in the middle Jurassic strata, on the western flank of the Aiguilles d'Arves ridge, is of completely different character to that of the Triassic gypsum – although generated during the same deformation phase. As mentioned above, they are consistently N–vergent and are truncated by the Eocene unconformity. **The folding occurs on two scales.**

1) Firstly, the middle Jurassic strata have been folded at a kilometric wavelength (this folding may be related to the growth of salt diapirs within the Triassic strata during extension). This undulation is recorded by the stratigraphic cut–offs of Jurassic bedding at the Eocene unconformity. Middle Jurassic beds are only preserved in the synclinal cores; elsewhere upper Lias shales are juxtaposed against the Eocene conglomerate.

2) Secondly, a smaller–scale N–vergent fold train observed at the Aiguilles d'Arves and Pointe des Chaudannes. The short limbs of these assymmetric folds are usually subvertical, but at Pte. des Chaudannes are overturned and dip at 55–65°S. Their fold axes fan reflecting the fan of the large scale folds and manifesting the internal strain of the Ultradauphinois thrust sheet.

4.4.2. The Aiguilles d'Arves–Pointe des Chaudannes Ridge.

This N–S trending ridge forms a prominent watershed between the Arvan and Valloirette Rivers. The ridge, which is composed of resistant Eocene sediments, comprises a series of peaks from the Aiguilles d'Arves (3513m) in the south to Pointes des Chaudannes (2521m) in the north (Fig.4.34). These peaks are accessible from the east via a series of gentle valleys, but from the west they present a formidable scarp (Fig.4.37). This morphology results from a regional eastward tilting of the Eocene strata.

The Eocene series lies with marked unconformity upon upper Lias and middle Jurassic sediments. The Eocene basal conglomerate that drapes the Mesozoic strata varies in thickness from 50–450m. The conglomerate consists of rounded clasts of basement gneisses and granites,

lower Liassic limestone, upper Liassic black shales and numerous quartz pebbles bound together by a sandy carbonate matrix. The rock is very resistant and so deformation tends to be accommodated by the overlying shales and flysch sediments.

The structures mapped within the flysch can be attributed to a single E–W compressive phase of post–Eocene age. During this phase however, a late E–vergent backfolding event has been superimposed upon an early W–vergent shearing and folding event. All the structures are colinear, their axes being subhorizontal and trending N–S.

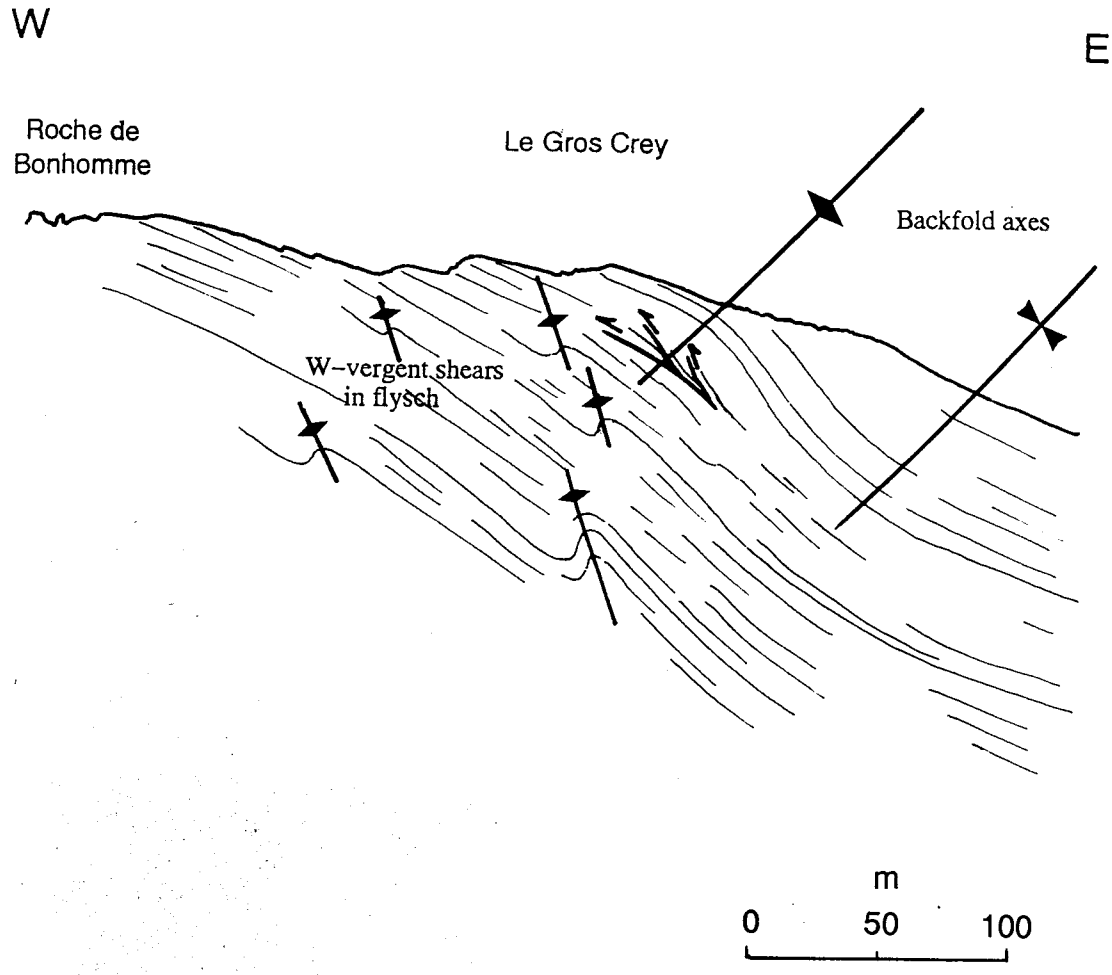
The Crete d'Argentiere, to the east of the Aiguilles d'Arves provides excellent exposures of west–vergent folds in the flysch, that are intimately related to numerous bedding–parallel thrusts. These thrusts are located in the shale layers of the flysch and displace the more competent sand layers, frequently ramping up–section towards the west and terminating in tip–folds. The simultaneous operation of numerous slip horizons leads to the folding and refolding of thrusts in a continuous shearing process (c.f. Fig.4.18). The short fold limbs are subvertical and strike SSE (N150°–160°E), thereby constraining the fold axial trend. The long limbs dip at 20°E, reflecting the regional dip of the Eocene unconformity.

On Le Crey Rond, east of Pointe de Pierre Fendue, the bedding strikes N001°–030°E and dips at 26–42°E. Mineral lineations in the shale units lie in the range N070°–128°E, but are mainly ESE–WNW directed. On Le Gros Crey, east of Roche du Bonhomme (Fig.4.41), small backfolds are well–developed. Vertical beds within this fold train strike N156°E.

Superimposed upon these shears and small scale E– and W–vergent folds are large scale flexures. Two such monoclines are mapped. Both are W–vergent and persistent along strike :

1) One outcrops close to the ridge crest and can be traced from the Aiguilles d'Arves to La Grande Chible (2932m). As one approaches the Aiguilles d'Arves the dip of bedding increases from 30°E to 60°E and one rapidly passes down–section through the flysch to silvery–

Fig.4.41 FIELD SKETCH OF THE EOCENE FLYSCH OUTCROPS, EAST OF ROCHE DE BONHOMME.
showing backfolded west-vergent shears



black shales, red shale and finally basal conglomerate. The main shearing clearly predates this flexure since west-vergent thrust ramps and asymmetric folds have been folded. The relationship between these phases is best exposed near the Roche du Bonhomme (2891m). At this locality the bedding dips at 25-35°E on either side of a zone of subvertical bedding which strikes N350°-354°E and small W-vergent folds are seen to swing around the backfold axis (Fig.4.41).

2) The other lies just west of the Briançonnais thrust front and can be traced from Bonnenuit to Valloire. West of Valloire, at Gorges de l'Ayguets and Gorges de l'Enter, the flysch bedding is flexed about a N-S striking, west-dipping axial plane. Bedding strike fans round from N356°E to N010°E and the dip rapidly increases from approximately 20°E to 80°E as one crosses the monoclinal axis. The bedding/cleavage relationships at all points on the monocline indicate W-vergent shear prior to flexural backfolding.

These large flexures affect the entire flysch series and may also affect the underlying Eocene conglomerate and Mesozoic succession. At the Aiguilles d'Arves the basal conglomerate is folded. The main shearing deformation, however, seems to be restricted to the hangingwall of a major shear detachment in the Eocene shales, above the basal conglomerate.

4.4.3. Summary of the Arvinche Deformation History.

1) The spectacular Arvinche folds in the Mesozoic strata are a result of two factors :

a) halokinesis of the Triassic strata during Mesozoic extension and

b) N-vergent compression that was demonstrably pre-Eocene and perhaps pre-Senonian (cf. Devoluy).

2) Considerable uplift was associated with this folding and major erosion occurred prior to the deposition of the Eocene series.

3) In post-Eocene times the region has undergone E-W compression.

4) The principal post-Eocene structure is the W-vergent Ultradauphinois Thrust and its splays.

5) Where Triassic evaporites are present, the Ultradauphinois Thrust sheet has glided upon the weak decollement with negligible strain being accommodated within the sheet itself.

6) Where Triassic evaporites are absent and the Ultradauphinois Zone has been thickened by imbrication and pure shear. This post-Eocene uplift is confirmed by uplift of the Eocene unconformity towards the south, its mean level rising 1km between Pointe des Chaudannes and the Aiguilles d'Arves (Fig.4.40). The development of thrust splays in the south and the concomitant strain gradient within the Ultradauphinois Thrust sheet also caused reorientation of the Arvinche fold axes.

7) Post-Eocene shearing is also manifest in the Eocene flysch series. At this level, the deformation sequence was :

a) early W-vergent simple shearing,

b) late E-vergent backfolding.

CHAPTER 5

INTERFERENCE OF ALPINE DEFORMATION PHASES IN PELVOUX, S.E.FRANCE.

In the vicinity of the Pelvoux massif three major phases of Alpine deformation have been superimposed upon Tethyan extensional geometries. The local expression of these phases was described in chapter 4. The aim of this chapter is to summarise the style of each phase and to compare their influence upon each other, their regional extent and intensity. The events are considered in three sections :-

1) **Triassic-Lower Cretaceous passive margin formation**. This phase, which was largely influenced by pre-existing Variscan lineaments, strongly controlled the subsequent phases of Alpine evolution.

2) **Pre-Nummulitic Alpine compression**. Due to the limited timing constraints available, this phase includes all Alpine deformation that is truncated by the unconformable Priabonian Nummulitic flysch series. As such, it comprises two distinct events - N-vergent deformation in the Arvinche (cf. Dévoluy) and WSW-vergent deformation in Pelvoux.

3) **Post-Nummulitic Alpine compression**. This phase comprises a system of WNW- to WSW-vergent basement thrusts that breach the Nummulitic flysch series.

5.1. MESOZOIC EXTENSION.

Rifting in the Triassic, the Hettangian-Sinemurian and particularly the Toarcian-Aalenian generated a series of tilted basement fault blocks on the Tethyan margin of SE.France. Such extensional geometries have been recognised in the external basement massifs, particularly in northern Pelvoux, e.g. the Bourg d'Oisans and Deux Alpes half-grabens in which a maximum of *ca.*4km of synrift sediments are preserved (Debelmas & Lemoine, 1970; Barféty et al, 1979; Lemoine, Gidon & Barféty, 1981; Gillcrist, 1988; Coward et al, 1991). Due to differential uplift

during the post–Nummulitic deformation phase, the structures of southern Pelvoux plunge at *ca.*30°S and are therefore less well–exposed than those of N.Pelvoux. However, two basement normal faults are recognised, the **Gondouins Fault** and the **Rouite Fault** (B.R.G.M. Orcières sheet, No.846 (1980)). A possible third fault, the **Fournel Fault**, is suggested by the Eocene subcrop pattern but does not outcrop. These faults control the basement topography and the distribution of synrift sediments (Fig.5.1).

1) The Gondouins Fault.

The Gondouins Faults, which downthrow to the east, form the western margin of the Cedera half–graben. Restoration of Alpine shortening reveals an original throw on the faults of approximately 2km (Fig.4.4). It is therefore of similar magnitude to the Ornon Fault (Barféty et al, 1979). An extensional offset is preserved today despite Alpine inversion, because although the Tourond Thrust is thought to have reactivated the Gondouin Fault at depth, it displays a breaching geometry at the surface (Fig.5.2). The Tourond Thrust propagated into the footwall of the early normal fault scarp (Gillcrist et al, 1987).

In the uplifted footwall of the fault only the Triassic and lower Liassic sedimentary cover is preserved. The localised development of a sandy unit at the base of the Triassic spilite suggests early activity on the Gondouins Fault. The lower calcareous Lias strata are equally developed in the hangingwall and footwall of the fault (Cedera and Tourond localities respectively), suggesting that major displacement on the fault did not occur until upper Liassic times. A thick sequence of upper Liassic sediments is preserved in the hangingwall of the fault (Fig.5.3). The eastward continuation of this graben is curtailed by the Cedera fold but it is suggested that it was once linked to the Morges, Méollion and Selle sub–basins.

2) The Rouite Fault.

The Rouite Fault downthrows to the south–southeast by approximately 140m (Figs.4.14 & 4.20). Its trend is normal to the main NW–SE Cevenoles trend of the Gondouins Fault and it acts as a transfer fault within the hangingwall strata of the latter. Displacement is inferred to pin towards the Gondouins Fault. Adjacent to the fault Triassic breccias outcrop

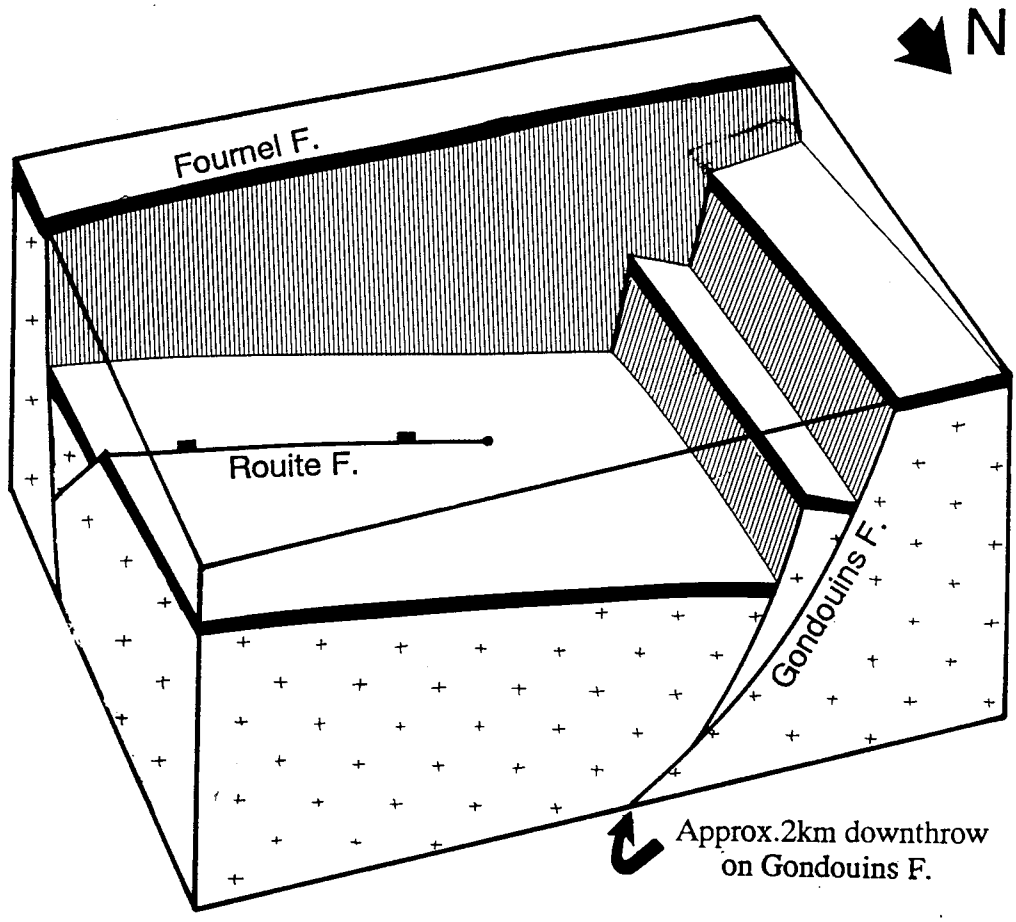
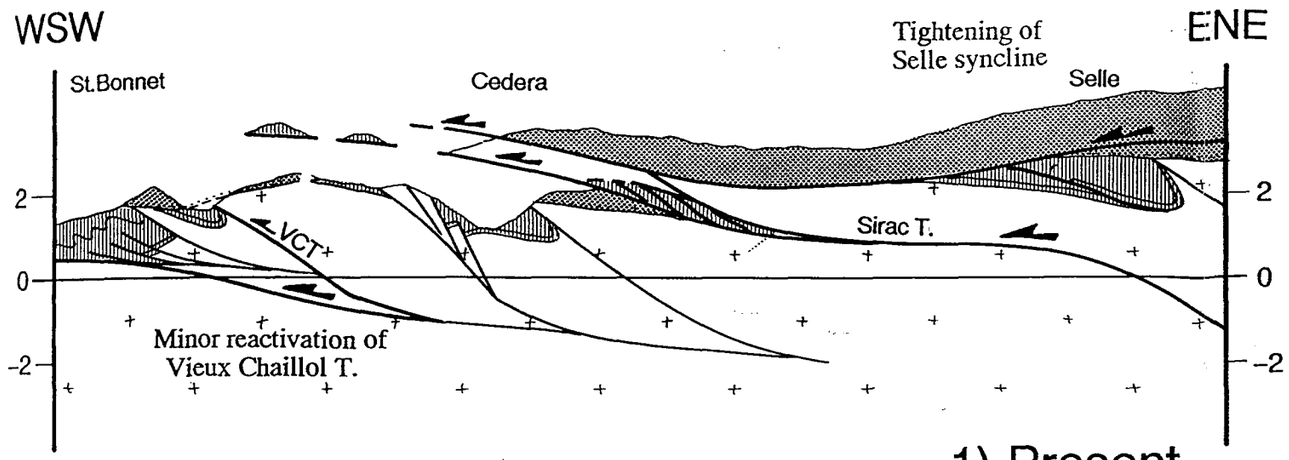


Fig.5.1. Block Diagram illustrating the Basement Geometry of the Cedera Half-graben after Liassic Rifting.

Crossed shading represents basement;
Solid shading represents Triassic pre-rift sediment

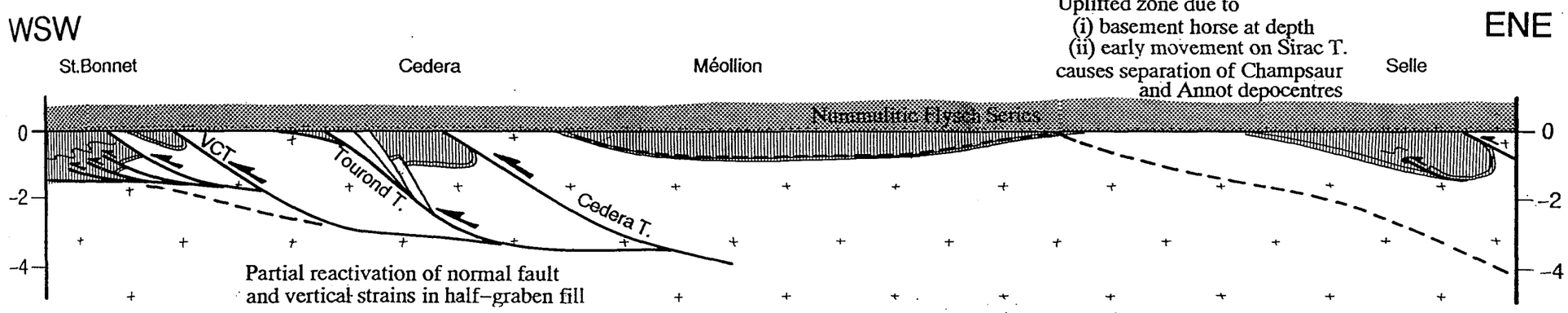
Fig.5.2. Phases of Structural Evolution in Southern Pelvoux.



15km shortening accommodated by Post–Nummulitic deformation
(Sirac T. system)

V=H

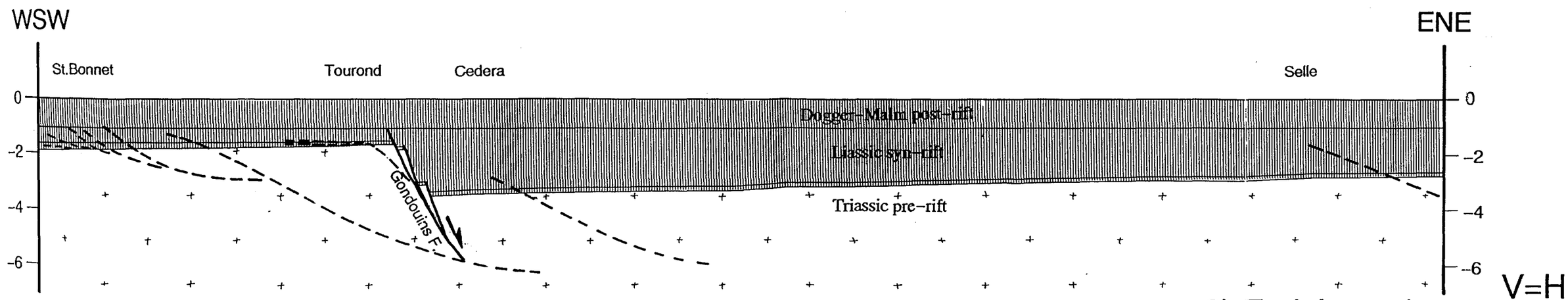
1) Present.



7–8km shortening accommodated
during Pre–Nummulitic deformation

V=H

2) End Eocene (flysch deposition).



V=H

3) End Jurassic.

suggesting early extensional activity, in common with the Gondouins Fault.

3) The Fournel Fault.

This fault is interpreted as an antithetic splay to a larger E–W trending fault that downthrows towards the north. The Fournel Fault, therefore trends normal to the Gondouins Fault and and is inferred to accommodate major displacement and form the southern boundary of the Cedera half–graben. The fault is not exposed but is inferred from field mapping data.

The main evidence for its existence is geometric. During both the Pre– and Post–Nummulitic phases of Alpine compression the southern margin of Champsaur has formed a lateral culmination. During the Pre–Nummulitic phase, the St.Bonnet, Vieux Chaillol and Tourond Thrusts all ramped laterally from basement into the Oxfordian "terres noires" shales. To the north of the lateral ramp uplift was sufficient to expose the basement during Nummulitic transgression. Again, in Post–Nummulitic times, uplift to the north of this lineament was much greater than to the south, as can be seen from the tilt of the Nummulitic unconformity. It is suggested that the persistence of this culmination throughout Alpine deformation is due to an inherited basement fault that has controlled basement uplift during both Pre– and Post–Nummulitic deformation phases.

It is possible that the absence of Mesozoic cover sediments in the Fournel window and the southeasterly tilting of the basement–cover unconformity at Dourmillouse may reflect Liassic uplift on the footwall shoulder of this bounding fault. This is illustrated in a NW–SE trending cross–section from Selle to Dourmillouse (Fig.4.26). In this interpretation the thick Triassic–Liassic succession at Selle represents an easterly continuation of the Cedera half–graben.

This basement topography could result from Pre–Nummulitic basement uplift due to thrusting. The trend of the Triassic cut–off between Fournel and Dourmillouse does not distinguish between the two models. Further stratigraphic analysis of the Liassic succession at Selle may distinguish between the models.

5.2. PRE-NUMMULITIC ALPINE DEFORMATION.

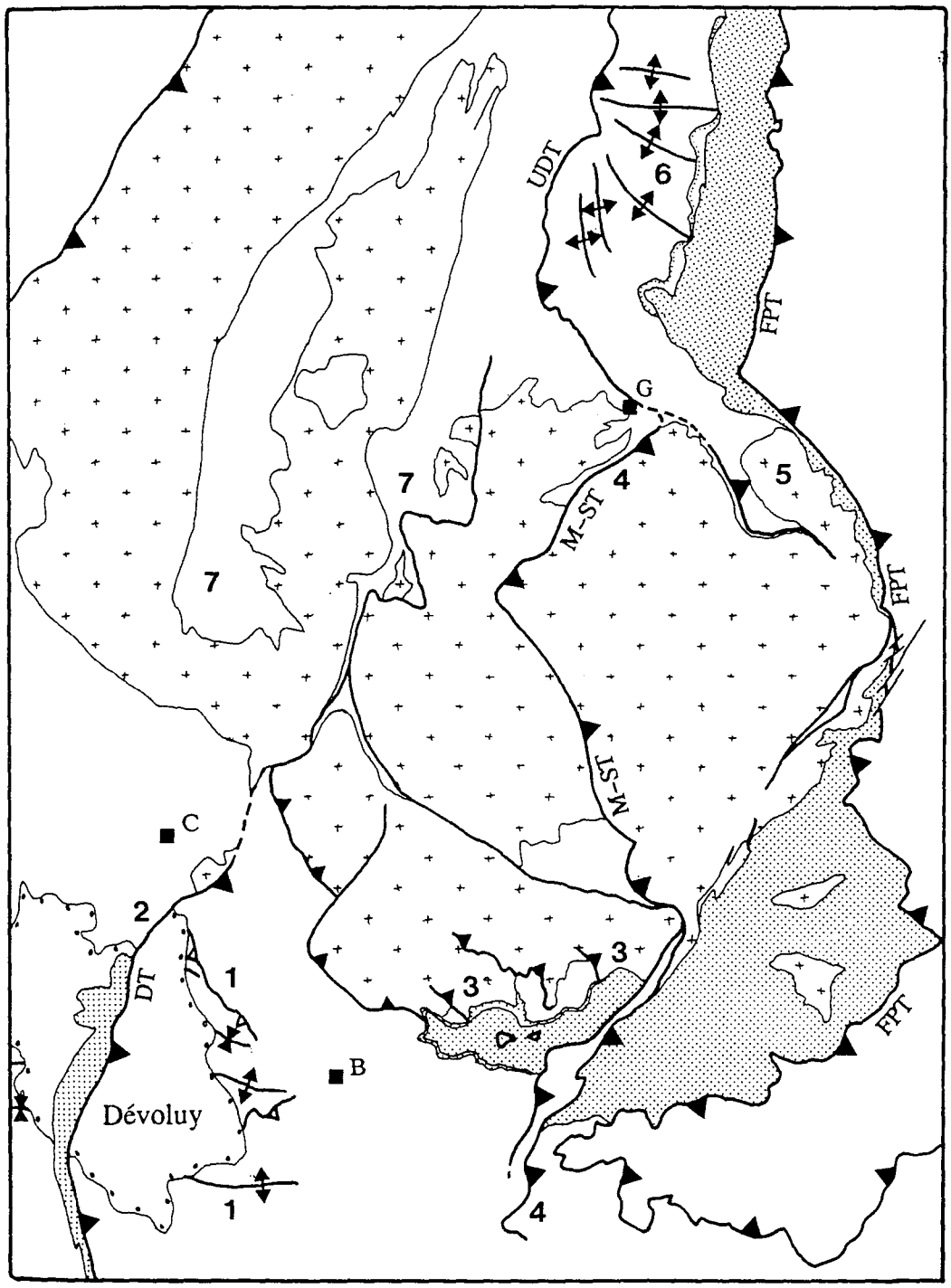
The Pre-Nummulitic deformation of the external zones of the French Alps has been recognised since the end of the last century (Lory, 1894). To this day, however, the regional extent and importance of this deformation remains poorly constrained. Our understanding of the "Pre-Nummulitic" phase is complicated by the fact that the Pre-Nummulitic phase of the Arvinche (Barbier, 1956; Barbier & Barféty, 1972; Barféty & Gidon, 1990) and that of Pelvoux (Lory, 1894; Gidon, 1954; Gidon, 1979) bear little similarity and are, in fact, different phases. The age of the Pre-Nummulitic phase in the Arvinche is constrained to the period middle Jurassic-late Eocene. It is more closely related to the N-vergent Pre-Senonian deformation of Dévoluy or Pyrénéan-Provençal deformation of the Alpes Maritimes than the WSW-vergent Pre-Nummulitic deformation of Pelvoux. Here the Pre-Nummulitic phase in the Arvinche and in Pelvoux will be regarded as two different events (Fig.5.3).

5.2.1. Pre-Nummulitic deformation in the Arvinche.

In the Arvinche this phase is characterised by E-W trending folds in the Mesozoic cover (Barbier & Barféty, 1972). The Jurassic cover is detached from the basement by the gypsiferous Triassic unit. The gypsum has flowed to produce diapiric walls that rise over 500m above their regional level at Génin. **Halokinesis is thought to have occurred during rifting in the late Liassic** (Graciansky, Rudkiewicz & Samec, 1986; Mascle et al, 1988). The overlying lower calcareous Lias sediments form an unbroken layer of uniform thickness. Diapirism, therefore, occurred during the upper Lias resulting in major N-S thickness variations in the synsedimentary U.Lias shales. Such diapirism has been observed in the Subbriançonnais Zone of Maurienne (Samec et al, 1988).

Further amplification and reorientation of these folds occurred during early Alpine compression because the middle Jurassic strata have been uplifted above the diapirs and

Fig.5.3. DEFORMATION PHASES IN THE PELVOUX MASSIF



GR, Grandes Rousses; FPT, Frontal Pennine Thrust; UDT, Ultradauphinois Thrust; M-ST, Meije-Sirac Thrust; DT, Dévoluy Thrust. B, St.Bonnet; C, Corps; G, La Grave. Large Arrowheads indicate Post-Nummulitic thrusts. Small arrowheads indicate Pre-Nummulitic thrusts. Open arrowheads indicate Pre-Senonian thrusts. Crossed pattern represents basement; blank shading represents Mesozoic cover; dotted shading represents Nummulitic flysch series.

1 - N-vergent cover folds truncated by Senonian unconformity.
2 - Post-Nummulitic Dévoluy thrust.
3 - WSW-vergent basement thrusts truncated by Nummulitic unconformity.
4 - Post-Nummulitic Meije-Sirac thrust system.
5 - Pre- and Post-Nummulitic movement at Combeynot.
6 - N-vergent cover folds truncated by Nummulitic unconformity (Pre-Senonian?).
7 - uncertain age of half-graben inversion in N.Pelvoux.

these high level undulations have been truncated by the Nummulitic unconformity. The middle Jurassic strata have also been folded about E–W trending axes. These folds are consistently N–vergent, irrespective of their position with respect to the long wavelength undulation axes. N–vergent basement thrusting in the Taillefer massif and the northern margin of the Pelvoux massif has been associated with this phase of deformation (Bartoli, Pécher & Vialon, 1974; Barféty & Gidon, 1990).

The extension of this Pre–Senonian or Pyrénéan–Provençal deformation phase to S.Pelvoux however, contrary to the interpretations of Barbier (1963) and Gidon (1979), is not substantiated in the field. However, if this phase comprised a high–level shearing of the upper Jurassic–Cretaceous succession with décolement within the "terres noires" facies, as it did in Dévoluy and possibly the Arvinche, then such evidence would have been eroded during Pre–Nummulitic uplift.

The age of these Pre–Nummulitic, N–vergent structures is poorly–constrained. They may belong to either the Pre–Senonian (Barbier & Barféty, 1972) or Pyrénéan–Provençal (Butler et al, 1986) phase, but, as mentioned earlier, it is unlikely that they are related to the Pre–Nummulitic phase of Pelvoux because transport directions are orthogonal. Distinction between the Pre–Senonian and Pyrénéan–Provençal phases is impossible in the field.

In addition to the N–vergent structures, **Barbier (1956) recognised W–vergent cover thrusts to the northwest of the Col du Lautaret that are truncated by the Aiguilles d'Arves flysch**. The origin of these structures was re–interpreted by Beach (1981). He argued that since the unconformity at the base of the flysch series has been sheared that these imbricates are of Post–Nummulitic age. This shearing, however, does not preclude the existence of Pre–Nummulitic imbrication in this area. In fact, in order to maintain lateral compatibility with the Pre–Nummulitic basement thrusts of Combeynot to the south it is inferred that Pre–Nummulitic E–W shortening did occur in the Arvinche (Trudgill, 1991).

5.2.2. Pre-Nummulitic Deformation in the Pelvoux Massif.

The structures that can be dated as Pre-Nummulitic in the Pelvoux massif have a totally different character from those just described and are therefore thought to belong to a different, and probably later, event.

In Pelvoux, the Pre-Nummulitic phase is characterised by thrust propagation folds (Ramsay & Huber, 1987) that are breached on either their short limb ("pli-faille") or long limb (Gidon, 1989) as the thrust-tip propagates through the fold. In Pelvoux both basement and cover rocks are involved in this style of deformation but it has also been described in the cover sediments of the Subalpine Ranges (Gidon, 1989). Due to erosion prior to the deposition of the Nummulitic flysch, the structures have been truncated such that cover sediments are only preserved in footwall synclines beneath the over-riding basement thrust sheets (Fig.5.2). Since the hangingwall cut-offs are not preserved the minimum estimate of shortening illustrated in chapter 4 (7-8km) may in no way reflect the actual shortening accommodated by this phase.

The minimum estimate places the eroded hangingwall cut-off of the breaching thrust as close to the footwall cut-off as possible. This method results in the modelling of the major structures of southern Pelvoux as large asymmetric folds that have been breached by minor late stage thrusts. In actual fact, the breaching thrusts may accommodate kilometric displacements. The importance of this phase of deformation is revealed by the uplift it generated. Between 1-2km of lower-middle Jurassic sediments are recorded in northern and southern Pelvoux and up to 3km of upper Jurassic "terres noires" sediments have been preserved to the southwest of Pelvoux (Kerckhove, 1979; Roux et al, 1988). Modelling of the Dauphiné basin suggests water-filled basin subsidence of approximately 2250m in Jurassic times (Rudkiewicz, 1988). A further 500-600m of lower Cretaceous and 600-700m of upper Cretaceous strata accumulated in Dévoluy. Approximately 6km of Mesozoic cover sediments accumulated in the Dauphiné basin. Although the Pelvoux massif remained relatively "high" throughout the rifting stage of Mesozoic extension (Roux et al, 1988) one can speculate that it subsided with the rest of

the margin during the upper Jurassic and Cretaceous periods, as the Ligurian Tethys opened. Between 3–4km of post–rift sediment may have accumulated on the Pelvoux massif. At the time of Nummulitic transgression, basement strata were subaerially exposed and thus **up to 4km of uplift may have occurred during Pre–Nummulitic deformation** (Fig.5.2). This exceeds the uplift generated during Post–Nummulitic deformation (*ca.*2km).

The Pre–Nummulitic phase of Alpine deformation was therefore of great importance in Pelvoux. In southern Pelvoux, where the Nummulitic flysch series is best preserved, it probably accounts for at least one third of Alpine shortening. This leads one to question whether certain undated structures of northern Pelvoux, which are generally assumed to be Post–Nummulitic are not in fact Pre–Nummulitic. Gillcrist (1988) suggested this possibility, although no incontrovertible evidence based on unconformable relationships is available. The criterion of structural style however suggests that the early inversion geometries of northern Pelvoux are related to the Pre–Nummulitic structures, and not the Post–Nummulitic breaching structures, of southern Pelvoux.

The Pre–Nummulitic structure of S.Pelvoux is illustrated in Figure 5.2. The Selle, Cedera and Muande structures all display the "pli–faille" geometry characteristic of the Pre–Nummulitic phase. The structures develop as tip–folds to basement thrusts that are subsequently breached at a late stage. At Tourond, the thrust geometry is different due to the presence of an inherited basement weakness in the form of the Gondouins Faults. It is possible that the normal fault plane is reactivated at depth, but the surface geometry indicates breaching of the footwall shoulder of the Gondouins basement fault block. To the west of the St.Bonnet thrust, deformation appears to affect the cover sequence only.

A feature of the Pre–Nummulitic thrust geometry is the exposure of basement between Selle and Méollion at the time of Nummulitic transgression. This region may have been a "high" during flysch deposition as well since the Gres du Champsaur and Gres d'Annot flysch sequences are of different provenance. The width of this high is unknown; the minimum width

possible is modelled in Figure 5.2.

As suggested above, the northward extension of Pre-Nummulitic deformation to the Arvinche is demanded by the fact that the Combeynot Thrust was active at that time. In accord with previous authors (Barbier, 1963; Beach, 1981; Gillcrist, 1988; Trudgill, 1991) the basement thrusts of Pelvoux are thought to ramp into the cover sequence in the Arvinche, producing imbricates of Triassic-Jurassic strata. If no expression of Pre-Nummulitic deformation occurs in the Arvinche (Beach, 1981) then the northern margin of the Combeynot massif must be interpreted as a steeply-dipping, E-W trending, transfer structure - perhaps a reactivated Jurassic fault (similar to the proposed Fournel Fault).

To the south of Pelvoux also, pre-Nummulitic folding has been described in the Ubaye valley to the west of Le Lauzet-Ubaye (B.R.G.M. Seyne sheet). At this locality pre-Senonian folds have experienced a second phase of growth prior to the Nummulitic transgression. This folding, as in southern Pelvoux, is oriented NW-SE suggesting SW-vergent compression - a direction that has remained constant in this region throughout Alpine times (Fry, 1989).

In the Pelvoux massif, where the Pre-Nummulitic deformation phase is best exposed, only the basement, pre-rift and syn-rift rocks are exposed. It is therefore strongly constrained by pre-existing basement geometries and offsets of the basement-cover contact generated during Liassic rifting. Deformation at this level is thick-skinned. At higher structural levels the deformation is predicted to adopt a thin-skinned style due to the drape character of the post-rift sequence. This change in structural style is attributed to the stratigraphic template and therefore can not be used to distinguish between the Alpine deformation phases. The erosion of the Nummulitic flysch series from the Subalpine chains, except for restricted outcrops, means that recognition of Pre-Nummulitic deformation in the post-rift sequences of these regions is impossible.

5.3. POST-NUMMULITIC ALPINE DEFORMATION.

This deformation phase comprises a set of three major, WNW- to WSW-vergent thrusts - the Frontal Pennine Thrust, the Ultradauphinois-Meije-Sirac Thrust and the Internal Belledonne-Dévoluy Thrust (Fig.5.3). Each of these structures breaches the Eo-Oligocene flysch series. In addition, Pre-Nummulitic structures such as those of the Combeynot, Selle and Muande localities were reactivated or breached and the inversion of the Deux Alpes and Bourg d'Oisans half-grabens may have occurred at this time.

The Pelvoux massif had been uplifted during both Pre-Nummulitic deformation and in the U.Eocene (Vernet, 1974) prior to displacement on the FPT, as is clear from the truncation of the Eocene flysch series at l'Yret in eastern Pelvoux. The massif, therefore, acted as a resistant block during the main Post-Nummulitic phase of Alpine thrusting. This has been emphasised by Bravard & Gidon (1979) who view Post-Nummulitic shearing in the Pelvoux massif as a predominantly strike-slip event. The data presented in chapter 4 confirms the major early Alpine uplift of Pelvoux but suggests a different style of Post-Nummulitic deformation. Many of the strike-slip faults of Bravard & Gidon (1974) have a major dip-slip component and these structures are interpreted as resulting from the collapse of the Briançonnais nappes into the underlying flysch series.

A pattern of Post-Nummulitic deformation in Pelvoux emerges in which zones where minor shortening is accommodated by the modification of pre-existing Mesozoic or Pre-Nummulitic structures are punctuated by low angle breaching thrust structures that accommodate major shortening. Thus, although all of the S.Pelvoux localities described in chapter 4 have suffered some Post-Nummulitic compression, major displacement is associated with the Sirac Thrust alone (Fig.5.2).

This is clear from Figure 5.2 which shows the Sirac Thrust ramping from basement to a flat within the Gres du Champsaur flysch series. Minimum displacement on the Sirac

Thrust is 5–6km but a further 11–12km is accommodated by the Méollion Thrust in its footwall. The erosion of this thrust system means that much greater displacements may have occurred but cannot be proven.

A feature of the Méollion Thrust sheet, which comprises the Triassic–U.Jurassic cover series, is the effect of the Cedera footwall ramp from the basement–cover interface to an intraflysch horizon. The competent Triassic and Lower calcareous Lias strata at the base of the Méollion Thrust sheet were imbricated by hangingwall collapse at the obstacle, whilst the incompetent shaley U.Lias–Malm sequence underwent ductile shear and is preserved as klippen to the south of Tourond and Muande.

The Vieux Chaillol Thrust (VCT) suffered minor reactivation during Post–Nummulitic times as is testified by the folding of the Nummulitic unconformity at Muande. Major displacement, however, was transferred to the basement–cover interface in the footwall to the VCT and St.Bonnet Thrust towards Dévoluy (Fig.5.3).

The flysch series in the hangingwall of the Sirac Thrust is strongly deformed – the deformation becoming more intense as one approaches the FPT in the east (Tricart, 1980). Shear–ing is concentrated in the in the Globigerina marl detachment near the base of the Nummulitic trilogy. Throughout SE. Pelvoux the flysch series itself is spectacularly folded into WSW–vergent chevron folds.

In the footwall of the Sirac Thrust, the Gres du Champsaur flysch is deformed in the same style except folding is NW–vergent. This change in vergence may be attributed to :–

(i) local stresses induced by impingement against the uplifted Pelvoux basement.

(ii) lateral pinning of deformation to the south of Pelvoux against the massif.

The dip of the Eocene unconformity in eastern and southern Pelvoux indicates marked Post–Nummulitic uplift of central Pelvoux. The culmination on the southern margin of

Pelvoux coincides with the Pre–Nummulitic lateral ramp culmination. This strongly suggests that both uplift phases were constrained by the Jurassic extensional structures.

To the northeast of Pelvoux, at Combeynot and Vallouise, the Post–Nummulitic deformation is characterised by backsteepening of the nummulitic unconformity. This monoclinal backfolding is also observed to the north of Pelvoux, along the Aiguilles d'Arves–Roche de Bonhomme ridge in the footwall to the Subbriançonnais Zone.

In the Arvinche, Post–Nummulitic deformation is accommodated principally by the UDT. Towards the south, the Triassic gypsum on which the UDT glides thins and is replaced by the dolomitic facies of Pelvoux. This results in the W–vergent imbrication of the Ultra–dauphinois Zone in the southern Arvinche (north of the Col du Lautaret) (Beach, 1981). These thrusts may have initiated in Pre–Nummulitic times as a lateral continuation of the Combeynot Thrust (Barbier, 1963), but the majority of these imbricates are interpreted as Post–Nummulitic and linked laterally to the north and south to the UDT and La Meije Thrust respectively.

Shortening estimates for the successive phases of Alpine deformation in Pelvoux are poorly constrained. Sections across southern Pelvoux indicate a minimum shortening of the Triassic pre–rift sequence of 7–8km, but this may represent only a fraction of the total Pre–Nummulitic shortening. Post–Nummulitic shortening was apparently greater, a minimum of 17km shortening being accommodated by the Sirac thrust system and approximately 40% shortening being observed in the flysch fold trains. Quantitatively the Post–Nummulitic phase appears to be the more important Alpine phase. The individual structures of this phase are of greater lateral extent and are found on a more regional scale. Pre–Nummulitic structures can be recognised wherever the Nummulitic flysch series is preserved and it is proposed therefore that they probably exist in northwestern Pelvoux and elsewhere in the external French Alps but are indistinguishable from the main Alpine post–Nummulitic structures.

CHAPTER 6.

AN INTRODUCTION TO THE HIMALAYAS.

The Himalayas, spanning some 3000km from Afghanistan to Burma and containing the majority of the world's highest peaks, represent the greatest mountain chain in the world. Together with the Tibetan Plateau, the relatively recent uplift of this region to an average elevation of approximately 5000m has formed a physical barrier that has strongly affected both the regional ecosystem and the global climate. Their geological importance as the prime example of a well-exposed, young collision belt has been recognised since the late nineteenth century. Much of the pioneering work was undertaken by British geologists of the Geological Survey of India. By the early twentieth century French, German and Italian workers became interested in this region, whilst Chinese geologists lay the foundations of Tibetan research. Today workers from around the world are adding to the body of Himalayan literature.

6.1. THE PLATE TECTONIC SCENARIO

The Himalayas have long been recognised (e.g. Argand, 1923) as the product of continent-continent collision between India and Eurasia. Their formation marks the culmination of a Wilson cycle that saw the birth and closure of the Neotethys Ocean. Intracontinental rifting in Permian times led to the progressive break-up of Gondwana during the Triassic and Jurassic periods. The spreading of the Lhasa, Helmand and Lut blocks from northern Gondwana in the early Triassic (Sengor, 1990), followed by the separation of India/Madagascar from Arabia/Africa in the upper Jurassic (Coster, Lawrence & Fortes, 1989) produced the marginal sag basins of the northwest Indian continent. Subsequent closure of Neotethys by subduction under southern Tibet during the upper Cretaceous and Palaeocene led to India-Eurasia collision in the early Eocene. Collision was preceded by ophiolite obduction - the timing of obduction being the subject of debate but occurring between the late Cretaceous (Fuchs, 1982; Searle, 1983; Searle et al, 1987) and the late Palaeocene (Garzanti, Baud & Mascle, 1987).

Initial collision was therefore between the passive Indian continental margin and the destructive Andean–type margin of southern Eurasia, the boundary being marked by the Indus–Tsangpo Suture Zone (ITSZ). Evidence of subduction is found along the Eurasian margin from Sumatra to Iran (Daly et al, 1991). The part of this margin involved in Himalayan collision is represented by the Lhasa block, where northward subduction of the Neotethys Ocean occurred throughout the Cretaceous period as is indicated by the calc–alkaline Transhimalayan, or Gangdese, batholiths (and associated volcanics) that intruded the Lhasa block at this time. Towards the west, Neotethys was subducted beneath the Ladakh and Kohistan island–arc complexes, which were separated by marginal basins from the Eurasian margin. These island–arcs were sutured to Eurasia during the latest Cretaceous along the Main Karakoram Thrust (MKT). The change from extensional to compressional arc conditions can be tied to the northerly migration of spreading in the Atlantic which initiated the slow clockwise rotation of Eurasia in the Upper Cretaceous. The western equivalent of the ITSZ is the Main Mantle Thrust (MMT) which lies to the south of the Kohistan/Ladakh complex.

Continued post–collisional compression has caused the indentation of India into Eurasia which has been accommodated by the northward propagation of shortening and thickening of the Tibetan crust (Cohen & Morgan, 1987), the southward propagation of imbrication of the Indian crust (Coward et al, 1988) and the eastward extrusion of southeast Asia (Tapponnier et al, 1982). Since 45Ma, approximately 2600km of convergence has occurred between India and Eurasia. Dewey et al (1988) suggest that almost 2000km has been accommodated by thickening of Asia – 1000km in Tibet and 980km north of the Tarim basin – whilst 440km was accommodated by shortening within the Himalayan thrust belt and a maximum of 200km by extrusion processes. This concurs with Coward & Butler (1985) who proposed 470km shortening within the Pakistan Himalayas but contradicts Tapponnier et al (1982) who favour 1000–2500km extrusion. Dewey, Cande & Pitman (1989) review the estimates of Dewey et al (1988), attributing up to 50% of the total India/Eurasia convergence to Himalayan thrusting.

Deformation of the Indian plate has been facilitated by two main processes. Firstly, as the Gangdise fore-arc impinged upon the Indian continental slope the thinned and extended Indian continental passive margin was inverted. Neotethyan marine conditions persisted throughout this phase, and although most of the evidence has since been eroded, several regressive pulses are recorded on the Zanskar shelf in the lower Eocene (Garzanti, Baud & Mascle, 1987). Such telescoping of the passive margin may have had little slowing effect upon the northward drift of India. However, Dewey et al (1988) report deceleration of India at 52Ma and 45Ma. The period prior to 45Ma may represent a phase of accommodation of shortening in Tibet by vertical pure shear and the inversion of the Indian passive margin accompanied by the initiation of Himalayan thrusting in Greater India. The second process of Indian plate deformation is thrust wedge propagation. This has continued episodically since collision to the present, during which time the Indian crust has suffered imbrication, metamorphism, uplift, erosion and extensional collapse.

6.2. ZONATION OF THE HIMALAYAS

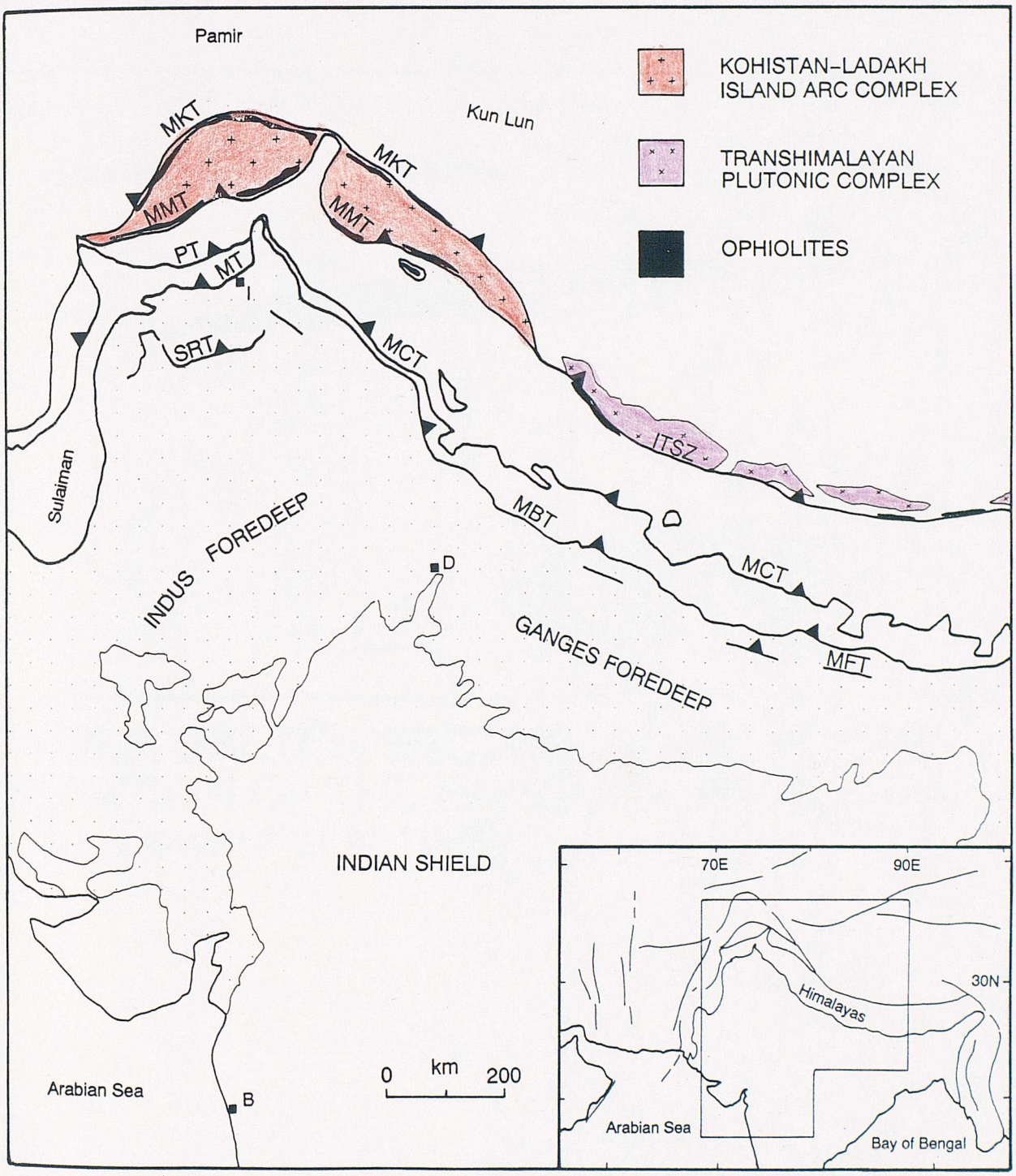
In the Main Himalayan arc, three principal zones have been identified in the footwall of the ITSZ (Figs.6.1 & 6.2) :-

1) The **High Himalayas,** which have been overthrust on the ITSZ by the Lhasa block, comprise a suite of metamorphosed Precambrian basement rocks and sedimentary supracrustals that have suffered Barrovian metamorphism, have been intruded by leucogranites and imbricated during the Himalayan orogeny.

2) The **Lesser Himalayas**, which have been overthrust on the Main Central Thrust (MCT) by the High Himalayas, comprise a suite that is relatively unaffected by Himalayan age metamorphism. Himalayan deformation is restricted to imbrication and folding of the Precambrian basement rocks of the Indian Shield and its mainly marine Phanerozoic cover sequence.

3) The **Sub-Himalayas**, which have been overthrust on the Main Boundary Thrust (MBT) by

FIG.6.1. THE ZONATION OF THE HIMALAYAS (after Gansser, 1991).



MKT, Main Karakoram Thrust; MMT, Main Mantle Thrust; ITSZ, Indus–Tsangpo Suture Zone; PT, Panjal Thrust; MCT, Main Central Thrust; MT, Murree Thrust; MBT, Main Boundary Thrust; SRT, Salt Range Thrust; MFT, Main Frontal Thrust. I, Islamabad; D, Delhi; B, Bombay

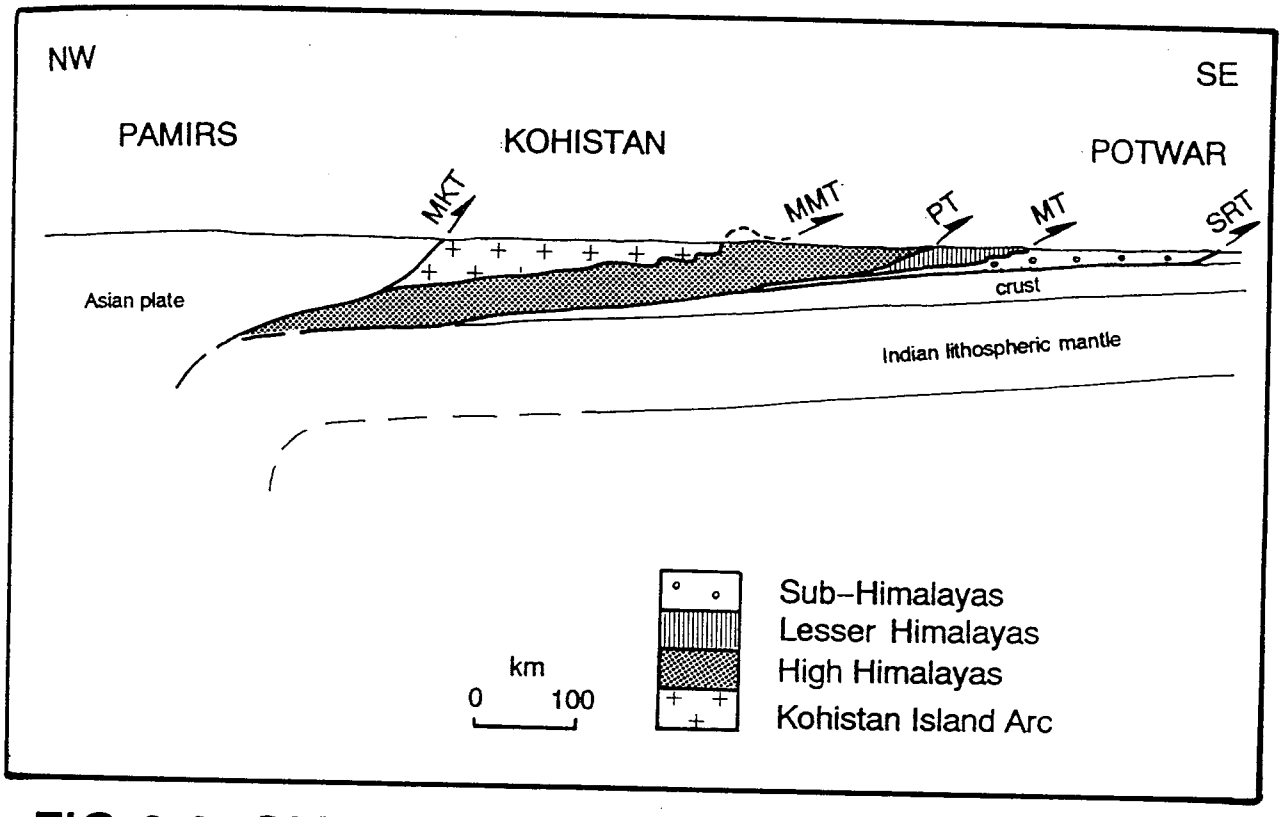


FIG.6.2. SKETCH SECTION SHOWING THE ZONATION OF THE PAKISTAN HIMALAYAS (after Coward et al,1988).

the Lesser Himalayas, are represented by foreland basin molasse sediments that have been deformed by the frontal structures of the Himalayan fold–and–thrust belt. The zone is bounded in the south by the Main Frontal Thrust (MFT) to the south of which undeformed foreland basin sediments are being deposited today.

In the Pakistan Himalayas these divisions become less clear. As mentioned above, the ITSZ can be traced westward as the Main Mantle Thrust (MMT) and hence the High Himalayas have been overthrust by the Kohistan Arc rather than the Lhasa block. The main difference however relates to the identification of the MCT to the west of the Hazara–Kashmir Syntaxis (HKS). In the Main Himalayas the MCT carried a major hot, crustal slab over the Lesser Himalayas resulting in an inverted metamorphic zonation due to downward heating. This is not observed in the Pakistan Himalayas where the deformation history of the internal zones is quite different (Treloar, 1989; Treloar, Williams & Coward, 1989). The inversion of the metamorphic gradient has been observed but this is attributed to post–metamorphic thrust imbrication placing higher grade rocks over lower grade rocks. This imbrication occurred after the peak of Himalayan metamorphism.

6.3. THE INTERNAL ZONES OF THE PAKISTAN HIMALAYAS.

As India was subducted under the Kohistan complex the upper–middle crust of the Indian Shield, in the footwall of the MMT, was introduced to higher temperature and pressure conditions. The early Himalayan deformation (D1) was the main fabric–forming event, imparting a mylonitic foliation to the Precambrian metamorphic basement and Phanerozoic cover sediments of the Indian Shield (Williams, 1990). Mineral lineations indicate SSE–vergent shearing and southwards from the MMT the style of deformation becomes less ductile. Continuing SSE–vergent shearing reworked the D1 fabric to produce D2 superposition structures (e.g. crenulation of D1, sheath folding). Treloar & Rex (1990) show that the metamorphic peak was reached by 45Ma, only 5–10Ma after collision. Further SSE–directed shearing since that time has imbricated this metamorphic pile into at least six major nappes that have been internally imbricated (D2a

phase). This post–metamorphic imbrication has produced a reverse metamorphic zonation by placing higher grade rocks over lower grade rocks. Rapid cooling of this nappe pile occurred between 24–15Ma suggesting that rapid uplift and unroofing took place at this time (Treloar & Rex, 1990), which coincides with the onset of significant molasse deposition. Imbrication of the internal zone must therefore have pre–dated or overlapped with the onset of uplift.

Late backfolding and backthrusting (D3) of the internal zone and extensional reactivation of the MMT has taken place during the later evolution of the Himalayan thrust wedge. In addition large scale cross–folding (D4), with fold axes perpendicular to the SSE–vergent thrust system, has been superimposed upon the region. This phase is held responsible for large scale uplifts such as the Besham Syntaxis (Williams, 1990).

Moving southwards one encounters successively lower grade rocks (metamorphic biotites are finally lost near the Panjal Thrust, north of Abbottabad) and one witnesses a transition from ductile to brittle deformation. The brittle–ductile transition occurs at the Oghi Shear (Coward et al , 1988) and is clearly indicated by the late Cambrian porphyritic Mansehra granite which is intruded into the Precambrian Tanawal Fn.. As one crosses the Oghi Shear the random orientation of the feldspar phenocrysts is lost and a preferred orientation acquired.

6.4. THE EXTERNAL ZONES OF THE PAKISTAN HIMALAYAS.

In the external zones, where Himalayan age metamorphism is of prehnite–pumpellyite grade or less and brittle deformation prevails, the crystalline basement of the Indian Shield is nowhere exposed suggesting decoupling at the basement–cover contact. The Phanerozoic sedimentary succession has been folded and imbricated since the middle Miocene, accommodating approximately 470km shortening (Coward & Butler, 1985).

Three distinct regions can be identified in the external zones of N. Pakistan :–

1) **the foothill zone of S.E. Hazara,**

2) **the North Potwar Deformation Zone (N.P.D.Z.)** and

3) **the South Potwar Zone.**

The structural style of each zone reflects the superposition of successive phases in the propagation history of the thrust wedge.

6.4.1. The Galiat.

Of these three regions the foothill zone, or Galiat, of S.E. Hazara has experienced the most complex deformation history. The rocks exposed in the zone have suffered three phases of deformation during the progressive SSE–vergent propagation of the Hazara/Potwar thrust wedge. In addition they are now being compressed by SW–vergent deformation related to the Kashmiri thrust wedge.

a) The first phase consisted of SSE–vergent imbrication of the MBT hangingwall. The structure is typified by imbricate thrusts that formed by footwall collapse, several such thrusts being traceable along strike into tip–folds.

b) The second phase involved the development of a major detachment, the MBT or Hazara Detachment. The hangingwall of this structure was translated at least 75km across a footwall flat located in the Rawalpindi Group molasse, as is manifest by the exposed MBT on the western limb of the HKS. The mechanism responsible for this scale of translation will be discussed later, but its main consequence is that the sediments of the Galiat and the northern Potwar, although presently juxtaposed, originally lay at least 100km apart. Stratigraphic variations between these two zones are therefore marked.

c) The third phase consisted of duplex formation in the footwall of the MBT. Initially the MBT acted as a SSE–vergent roofthrust to this duplex, but later the duplex propagated to the south of the MBT and the duplex adopted a "passive–roof" geometry. Deformation in the hangingwall of the roof–backthrust is manifest as **back–folding at the surface** in the Galiat and the N.P.D.Z..

d) In recent times the Galiat has suffered NE–SW compression. This is expressed in

two ways. Since the strata of the Galiat were already sub-vertical at this time, reactivation of pre-existing thrust- or bedding-planes was rendered difficult. Instead the Hazara region has suffered a widespread crumpling or buckling in which NNW- or SSE-plunging fold axes have been superimposed upon the earlier sub-horizontal, ENE-trending backfolds. The second manifestation of NE-SW compression is the regional uplift of eastern Hazara which is regarded as large-scale tipfolding above the blind prolongation of the NW-SE trending, SW-vergent Kotli Thrust.

6.4.2 The North Potwar Deformation Zone.

The NPDZ is bounded in the north by the topographic break that separates the Potwar Plateau (average altitude = 500m) from the Galiat foothill zone (average altitude = 2000m). This mountain front corresponds to the southern outcrop limit of the resistant carbonate sequences. The NPDZ itself is characterised, at the surface, by exposures of poorly-consolidated and steeply-dipping Miocene molasse. The exposure of Miocene molasse implies that the NPDZ has suffered considerable uplift that caused the erosion of the entire Siwalik Group succession. The southern limit of this zone of uplift is marked by the northern limb of the Soan syncline. Magnetostratigraphic dating has shown that the 2Ma old, steeply-dipping beds of the Dhok Pathan Fn. are draped by 1.9Ma old, sub-horizontal beds of the Lei conglomerate (Johnson et al, 1979; Johnson, Raynolds & Burbank, 1987). The age of the last phase of uplift in the NPDZ can therefore be tightly constrained at approximately 2Ma.

Despite strong deformation only Rawalpindi Group molasse sediments are exposed throughout the NPDZ. This implies that these Miocene strata have been decoupled from the underlying shelf sequence. The decoupling horizon lies within the over-pressured Murree Formation. The style of deformation that is observed above the detachment at the surface is in great contrast to that inferred from seismic data below the detachment.

At the surface, bedding has been imbricated and folded by SSE-vergent deformation but then suffered from the NNW-vergent backfolding phase that produced the sub-vertical bedding

dips of the NPDZ. At depth, the Cambrian–Eocene sequence has been imbricated to form a duplex – the individual thrust horses being observed as anticlinal crests on unpublished commercial reflection seismic data that have been uplifted from a regional elevation of approximately 3.7–3.8 seconds TWT to between 2.3–3.1 seconds TWT. Locally the imbricates rise to 0.7 seconds TWT and even breach the surface in the west at Khairi Murat. These thrusts are consistently SE–vergent and floor near the basement–cover unconformity.

The front of the duplex is blind and represented by a triangle zone that transfers SSE–vergent forward displacement onto the NNW–vergent roofthrust. This triangle zone geometry is responsible for the monoclinal shape of the northern limb of the Soan syncline. Note that the Riwat thrust that forms the southern limb of the Soan syncline is an earlier structure (3.7Ma). Hence the Soan syncline originated as a broad, flat–bottomed syncline that has been gradually tightened by the addition of successive imbricates to the N. Potwar duplex. The south–south–eastward propagation of the duplex caused the uplift and subsequent erosion of the molasse material in its hangingwall, producing the progressively broadening NPDZ. At least 2.5km of upper Miocene–Pliocene molasse has been eroded from the NPDZ as a result of this uplift.

6.4.3 The Southeastern Potwar Plateau.

Deformation in the NPDZ ceased in early Pliocene times (~5–4Ma) and since that time Himalayan shortening has been active to the south of the Riwat thrust in the southern Potwar. The Potwar Plateau is bounded in the south by the Salt Range which formed as the allochthonous Potwar Plateau–Salt Range thrust sheet that ramped over a basement normal fault to become emergent. Towards the east the SRT changes polarity and continues along strike as the Mahesian backthrust. Displacement along the SRT/Mahesian structure diminishes towards the east and shortening is accommodated by a series of structures. Therefore, the shortening accommodated by the SRT in the west is distributed across seven distinct structures in the SE. Potwar.

This change in structural style between the southwestern and southeastern Potwar has been

attributed to a number of criteria, all of which suggest that extra thickening of the thrust wedge was required towards the east for propagation of the wedge on its basal detachment to continue (Davis, Suppe & Dahlen, 1983). Butler et al (1987) suggest that the Potwar Detachment ramps laterally towards the east from within the Salt Range Formation evaporites to the Tertiary molasse series. The loss of the easy-slip salt horizon in the east is deemed to account for thickening of the wedge. Although ramping may occur further to the east, seismic data shows that in the eastern Potwar the Potwar Detachment remains in the Salt Range Formation (Pennock et al, 1989).

Jaume & Lillie (1988) attribute the buckling of the eastern Potwar to an eastward decrease in basement dip across the Potwar Plateau. Along a NNW-SSE trending profile across the eastern Potwar the basement dip is less than 1 degree towards the NNW. Since the topographic slope is also small, the wedge taper is negligible and wedge propagation could not continue without internal wedge thickening.

Descriptions of the structural evolution of the Potwar Plateau have, in general, overlooked the truncation of the Mesozoic carbonate sequence by the base Palaeocene unconformity. The effect of this unconformity is to reduce the thickness of the Mesozoic wedge from approximately 1.5-2km in the Trans-Indus Ranges (Leathers, 1987) to only 100-200m in the eastern Salt Range. Hence, in the west, structural style is dominated by the competent carbonate shelf sequence whilst towards the east, the incompetent molasse strata dictate the structural style. Hence, from west to east across the Potwar, the single major SRT sheet is replaced by numerous pop-ups, box-folds and low angle thrusts.

6.4.4 The Hazara-Kashmir Syntaxis.

A principal feature of northern Pakistan is the structural re-entrant or half-window called the Hazara-Kashmir Syntaxis. It has been documented since the last century and described in some detail by Wadia (1932). More recent work has been hampered by the political division

between India and Pakistan which makes access to some regions difficult. Various factors have contributed to the evolution of the HKS, but these will be discussed later. Of major importance however is the role of tectonic transport direction variation and the thrust interference that it causes. In Hazara, on the western limb of the HKS the Hazara thrust wedge is propagating towards the SSE (Yeats & Lawrence, 1984; Coward et al, 1988). Concurrently, in Kashmir, on the eastern limb of the HKS, the Kashmir thrust wedge (the northwestern limit of the Main Himalayas) is propagating southwestwards (Burbank, 1983; Burbank, Raynolds & Johnson, 1986). The primary origin of the syntaxis is not, therefore, the indentation of a resistant basement block nor rotation of a linear mountain chain due to temporal variations in transport direction. Rather, it is a product of thrust interference and extra thickening that has created a domed-shaped zone of uplift which, when eroded, forms a half-window.

This interference is displayed in two main ways. The presence of pre-existing structures related to thrust wedge "A" may constrain the propagation of later structures belonging to thrust wedge "B", e.g. the Pabbi fold. Alternatively, structures related to thrust wedge "A" may be superimposed upon those of thrust wedge "B", e.g. the Ayubia cross-folds.

The presence of two co-existing perpendicular transport directions in the HKS demands a three-dimensional approach to regional structural interpretation. In this area the plane-strain assumption of cross-section balancing is often invalid. This has implications at all scales from outcrop mapping to foreland basin modelling.

6.5. STRATIGRAPHIC EVOLUTION OF THE EXTERNAL ZONES

Stratigraphic controls have strongly influenced the style of structural deformation in the external zones. The Palaeozoic evolution of Gondwana, its break-up during the early Mesozoic with the formation of the Indian continental margins and eventual Eocene collision provide the foundation on which Pakistani stratigraphy stands. The rheology of the stratigraphic units, and their regional extent, has determined much of the structure observed today. Hence, the contrast

between the Sub-Himalayas of the Potwar Plateau and the Main Himalayan arc can be related to the distribution of Eocambrian salt. Likewise, the contrast in structural style between the western and eastern Potwar Plateau can be related to the development of the base Palaeocene unconformity.

6.5.1. The "Wilson" Cycle : Gondwana Break-up to Collision

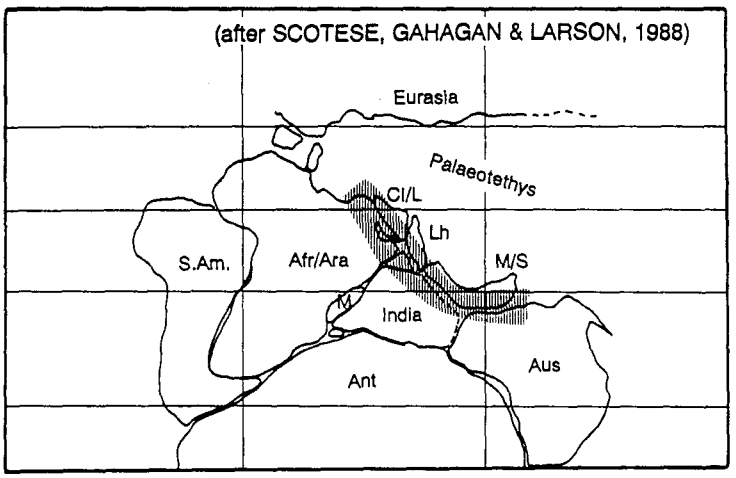
The conventional fit of the southern continents to produce Palaeozoic Gondwanaland places Madagascar adjacent to Somalia (Smith & Hallam, 1970; Norton & Schlater, 1979; Scotese et al, 1988), rather than Mozambique (Barron, 1981); closes the Red Sea and Gulf of Aden and then fits the western margin of India and the Seychelles to Madagascar. This results in the approximate juxtaposition of Pakistan and Oman (Fig.6.4). The locations of the Lut, Helmand and Lhasa blocks is more subjective but palaeontological and palaeomagnetic studies prove that these microplates are of Gondwana affinity. Most Gondwana reconstructions place the Helmand block either between or just north of Pakistan and Oman (Mountain & Prell, 1990; Sengor, 1990). A close correlation between the Palaeozoic and early Mesozoic successions of Pakistan and Arabia should therefore be possible.

The continental passive margins of India have different ages due to the progressive break-up of Gondwana during the early Mesozoic (Fig.6.3). In Pakistan the effects of two passive margin forming events are to be expected. Firstly, rifting of the Lut/Helmand/Lhasa blocks from Gondwana during the Permian (Béchennec et al, 1988) and opening of the Neotethys Ocean in the early Triassic (at approx. 250-240Ma) produced the northern margin of India, represented today by Zanskar (Searle, 1986; Garzanti et al, 1987). This margin has since been strongly deformed during the Himalayan collision. Secondly, the rifting of India/Madagascar from Africa/Arabia during the Triassic to middle Jurassic followed by oceanic spreading in the Somali and Mozambique basins in the upper Jurassic (from approx. 157Ma) produced the western margin of India (Coster et al, 1989).

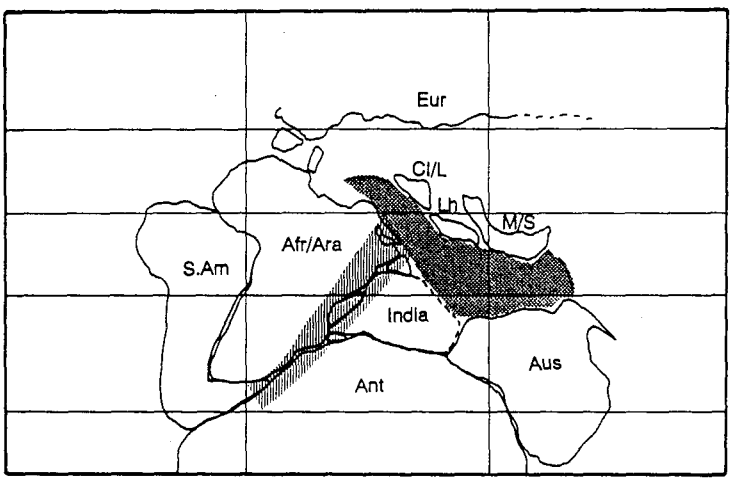
The Mesozoic rifting of Gondwana followed pre-existing Precambrian trends (Fig.6.4).

FIG.6.3. GONDWANA BREAK-UP AND INDIA'S PASSIVE MARGIN DEVELOPMENT.

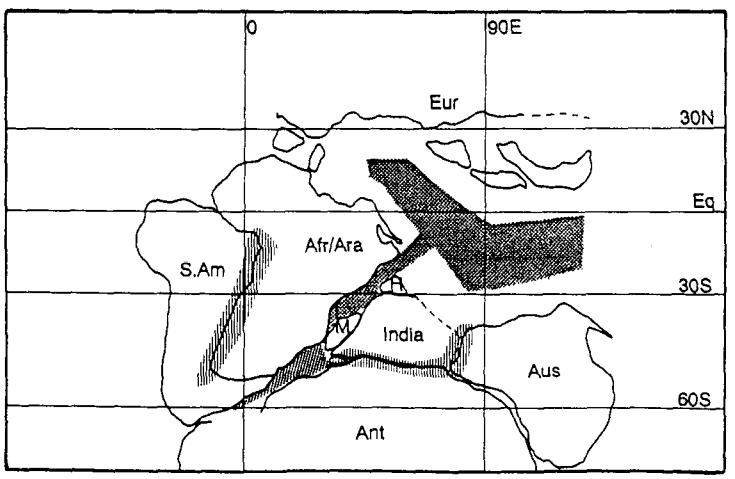
1) PERMIAN-EARLY TRIASSIC



2) MID TRIASSIC-MID JURASSIC



3) UPPER JURASSIC-EARLY CRETACEOUS



Active Rifting Active Spreading

Afr/Ara, Africa-Arabia; S.Am., S. America; Aus, Australia; Ant, Antarctica;
M, Madagascar; H, Helmand; Cl/L, Central Iran-Lut; Lh, Lhasa; M/S, Malaysia-Sumatra.

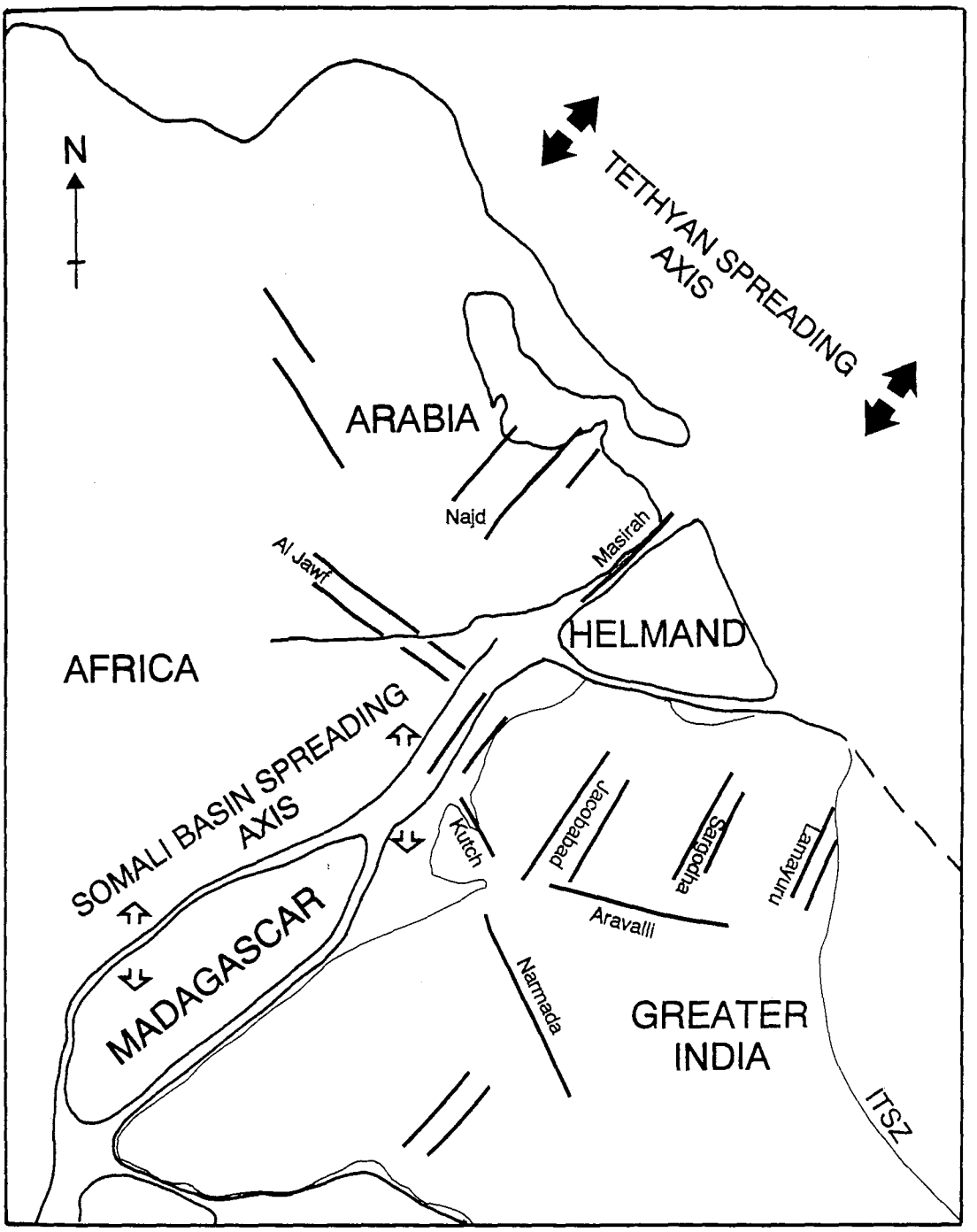


FIG.6.4. PRECAMBRIAN LINEAMENT TRENDS OF GONDWANA

The northern passive margin, although much deformed, was orthogonal to the main Aravalli trend of the Indian Shield. The extension of this margin can be observed on the northeastern passive margin of Arabia such as the Hawasina basin of Oman and the Iranian portion of the Arabian plate (Béchennec et al, 1988). The trends exposed in the basement outcrops of western Saudi Arabia and Yemen e.g. Al Jawf graben, correlate well with the Aravalli trend on Permian reconstructions of Gondwana. Late Triassic to middle Jurassic rifting of East and West Gondwana to produce the Pakistani passive margin of the Indian plate reactivated the Najd trend of the Arabian basement. This trend corresponds to the main basement structures of the western Indian margin which compartmentalise the principal depocentres of Indus basin, such as the Jacobabad and the Sarghoda–Shahpur "highs" (Fig.6.4).

India finally became isolated when it rifted from Madagascar. During the Cretaceous and Palaeocene, India's marginal sag basins entered their passive subsidence phase. Stratigraphic variations are principally a response to eustatic sea–level variations, sediment supply rates and asymptotically decreasing subsidence rates. The stratigraphic succession of Pakistan documents major transgression in Cenomanian times that re–introduced shelf carbonate deposition (Kawagarh Fn.) to the margin after Aptian–Albian shallow marine to continental clastic sedimentation (Lumshiwal–Giumal Fn.). This can be attributed to worldwide sealevel rise at this time associated with widespread oceanic spreading. Towards the end of this period sedimentation was interrupted by ophiolite obduction and finally collision in the early Eocene. Marine conditions finally departed in late middle Eocene times.

6.5.2. The Base Palaeocene Unconformity

The Pakistan passive margin was influenced in late Cretaceous times by pronounced regression which resulted in sub–aerial erosion of the entire shelf, the coastline being near the shelf–slope break. This regression is represented by the Stumpata and Hangu Formations and marks the most important unconformity in Pakistan. Near the feather–edge of the passive margin, in the eastern Salt Range, coal–bearing Hangu Fn. overlies Cambrian strata (Shah, 1977) – the entire Mesozoic sequence having been eroded. On the outer shelf, preserved in Zanskar,

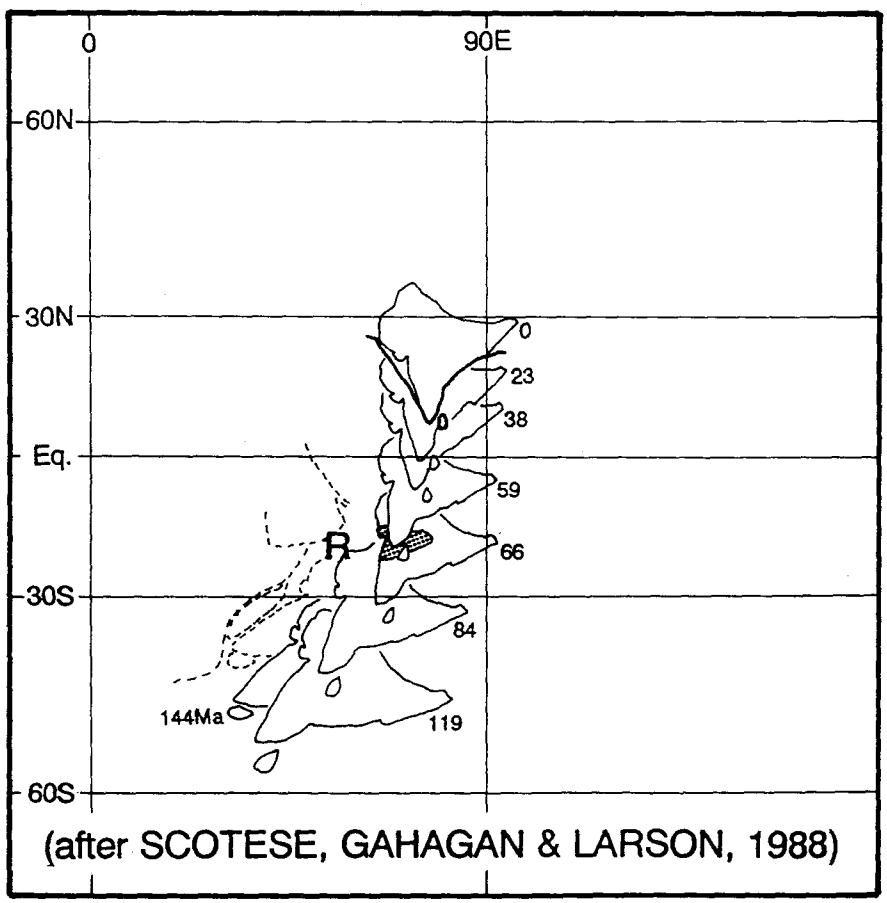
upper Cretaceous carbonates are overlain by the terrigenous Stumpata conglomerate (Garzanti et al, 1987). The lower Palaeocene is marked by major transgression that shifted the coastline at least 700km onto the Indian Shield, causing a return to carbonate shelf conditions (Lockhart Fn./Dibling lst.) (Shah, 1977; Gee, 1981; Garzanti et al, 1987).

Various causal mechanisms have been proposed for this regressive–transgressive cycle. Certain authors (Garzanti, Baud & Mascle, 1987) imply a eustatic origin, but such a dramatic sea–level fluctuation does not fit well with the Haq sea–level curve (Haq, Hardenbol & Vail, 1987). The scale of this regressive–transgressive cycle demands that regression was caused by positive uplift that produced approximately 1000m erosion and which was followed rapidly by a return to shallow marine carbonate conditions. Searle (1986) suggested that ophiolite obduction onto the northern margin of the Indian plate in late Cretaceous times would be associated with crustal foundering and thereby produce the Lower Palaeocene transgression. In this model the uplift that caused regression and subaerial exposure of the entire Indian shelf is linked to compression during plate kinematic changes that culminated in widespread ophiolite obduction along the southern margin of Neotethys. Garzanti, Baud & Mascle (1987) argue that obduction in the ITSZ occurred much later and therefore could not have caused the basal Palaeocene unconformity.

Such a regressive–transgressive cycle can be modelled by the drift of India across the Reunion mantle hotspot in the late Cretaceous (Fig.6.5). This model can produce both the required amount of uplift and the required uplift rate – factors which can not be accounted for by either the eustacy or obduction models.

Numerous plate kinematic reconstructions of the Indian Ocean have been presented in the last decade (Johnson, Powell & Veevers, 1976; Morgan, 1981; Owen, 1983; Patriat & Achache, 1984; Scotese, Gahagan & Larson, 1988; Powell, Roots & Veevers, 1988). They are based upon the integration of three main criteria. Firstly, oceanic spreading histories are constrained by seafloor palaeomagnetic reversal data. Secondly, the identification of the continent–

FIG.6.5. THE DRIFT OF INDIA AND THE REUNION HOTSPOT : CONTROL ON THE BASAL PALAEOCENE UNCONFORMITY.



Dashed outline represents Gondwana fit at Dogger/Malm boundary (156Ma). Dotted area represents eruption of Deccan Traps. Note that longitudinal errors will be greater than latitudinal ones in the reconstructions. In Late Cretaceous times NW India drifts past the Reunion hotspot

ocean transition and passive margin analysis constrains the shape of the continental margins prior to oceanic spreading. Lastly, palaeomagnetic data from the continents can constrain the latitudinal position of the continents through time with respect to the magnetic pole. If one assumes that mantle hotspots are relatively static (Morgan, 1981; Duncan & Richards, 1991) then when one superimposes the present day position of the Reunion hotspot upon these reconstructions it is clear that during the late Cretaceous India was drifting rapidly northwards and crossed over the hotspot. The passage of India across this thermal anomaly caused rapid uplift as predicted by White & McKenzie (1989) but since India was moving at approximately 15cm/yr it soon moved off the hotspot and began to subside once again. Uplift and doming above the Reunion hotspot culminated in the rifting of the Seychelles/Mascarene Plateau from India and the outpouring of vast quantities of continental flood basalts, the Deccan Traps, at 65Ma (Cox, 1989). The Deccan region is underlain by less dense, depleted lithospheric mantle and therefore has remained uplifted to this day despite the passage of India away from the hotspot. This remnant uplift of 1000–2000m is characteristic of all flood basalt provinces (White & McKenzie, 1989; Cox, 1989).

6.5.3. Continent–Continent Collision.

The Mesozoic evolution of Pakistan was one of passive margin development; an initial rift-dominated phase being followed by a thermal subsidence phase. Pakistani stratigraphy displays several major transgressive–regressive cycles that reflect eustatic variations and hotspot-related uplift. Passive margin subsidence, however, was brought to a close by collision with the "Andean-type" margin of southern Asia in early Eocene times (55–50Ma).

"Collision" is here defined as the time at which the Indian continental slope impinged upon the Gangdese forearc. The exact timing of collision is still a subject of debate, however most estimates lie between 55–45Ma. Part of the uncertainty stems from the diachroneity of collision from west to east along the Himalayan arc, collision having occurred first in the vicinity of the Northwest Syntaxis and progressively migrated eastward. However, a variety of criteria is used to constrain collision. These include :–

1) The end of carbonate deposition on the Indian passive margin,

2) The onset of inversion of passive margin extensional structures,

3) The transition from marine to continental deposition on the Indian passive margin,

4) The end of melange sedimentation,

5) Dating of the youngest clasts within the melange,

6) The cessation of calc–alkaline magmatism in the Gangdese arc,

7) The peak of Himalayan metamorphism in the Indian plate,

8) The decrease of spreading rates and reorganisation of spreading ridges in the Indian Ocean and

9) The marked deceleration of the northward drift of the Indian continent (Fig.6.6).

No single criterion will accurately date collision as defined above. The cessation of carbonate shelf conditions and calc–alkaline magmatism and the onset of molasse sedimentation may occur some time after "collision", whilst melange clasts may considerably predate collision. Himalayan metamorphism, Indian Ocean events and the arrestation of India may not commence until long after collision. The initial impingement of India upon Asia could be dated by inversion events on the Indian passive margin – but this evidence is often disguised by subsequent deformation.

Garzanti, Baud & Mascle (1987) date the onlap of calc–alkaline volcaniclastics (Chulung La Fn.) across the shelf carbonates (Kesi Fn.) of the Zanskar passive margin at 54Ma. This represents the first northerly terrigenous sediment source and therefore the proximity of the Gangdese arc. To the south, on the Punjab Platform, carbonate environments persisted into the middle Eocene indicating that, even after collision, a restricted Neotethys survived. Searle et al (1989) quote a youngest age of 37Ma for Ladakh plutonism (K–Ar date). This certainly post-dates the end of oceanic subduction.

Indian Ocean data record a major plate reorganisation between Anomalies 22–20 at which time spreading between India and Australia ceased, India–Australia fused to become one plate

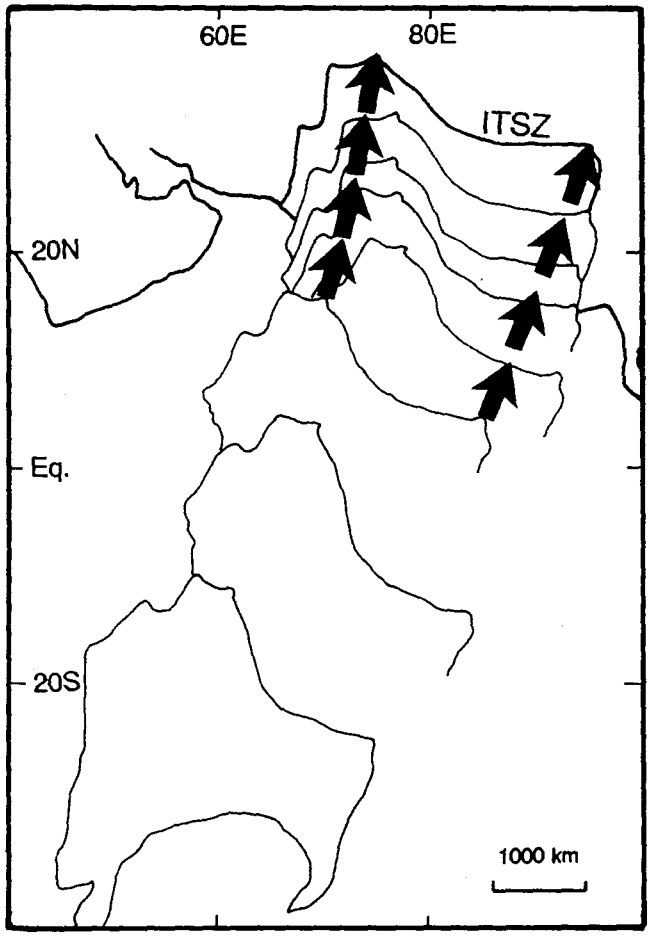


FIG.6.6. MOTION OF INDIA SINCE 70Ma, SHOWING DECELERATION AND CONTINUED CONVERGENCE SINCE COLLISION (after Patriat & Achache, 1984).

Arrows indicate displacement of the suture zone since collision.

and spreading between Australia and Antarctica accelerated. This was a direct response to India–Asia collision (Patriat & Achache, 1984; Scotese, Gahagan & Larson, 1988; Besse & Courtillot, 1988). **Initial collision occurred therefore in the early Eocene at approximately 54–52Ma.**

6.5.4. Post–Collisional Convergence.

Dewey et al (1988) indicate distinct changes in India's velocity at 52Ma (140–95mm/yr) and 45Ma (95–71mm/yr) and a gradual slow down to *c.*50mm/yr since then, suggesting that initial collision occurred at approximately 52Ma and that by 45Ma buoyancy forces were sufficient to begin "indentor" tectonics. Treloar & Rex (1990) show that the peak of Himalayan metamorphism on the Indian plate was reached by approximately 45Ma. This data suggests that between 50–45Ma the Indian crust was subducted beneath Tibet, causing a rapid increase in P–T conditions, until buoyancy forces equalled the slab–pull force. Thereafter India acted as an indentor; a vertical pure shear deformation propagating into Tibet (Dewey et al, 1988; Dewey, Cande & Pitman, 1989) whilst a simple shear deformation affected the rocks of the Indian plate in the footwall of the MMT–ITSZ (Treloar, Williams & Coward, 1989).

After the onset of indentor tectonics major shortening took place in Tibet and the Tien–shan, ongoing convergence causing the deformation front to propagate away from the collision boundary (Cohen & Morgan, 1987), whilst Himalayan shortening continued south of the suture. Obtaining accurate estimates for the partitioning of shortening is a topic of active research, but currently available data suggests that 50–66% of the total India/Eurasia convergence was accommodated north of the ITSZ, the proportion increasing towards the east (Dewey, Cande & Pitman, 1989). This implies that approximately 1000km shortening has occurred south of the ITSZ in the Himalayas, a figure twice that of previous estimates (Coward & Butler, 1985; Dewey et al, 1988).

Estimates of Himalayan shortening are necessarily poorly constrained. The author's research demonstrates at least 200km shortening in the footwall of the Nathia Gali Thrust (NGT), suggesting that external zone deformation accounts for approximately 250km shortening. When

shortening in the internal zones and on the Zanskar shelf is added to this figure, Coward & Butler's (1985) estimate of 470km Himalayan shortening appears conservative.

6.6. THE STRATIGRAPHY OF THE GALIAT AND POTWAR REGIONS OF PAKISTAN.

Extensive stratigraphic research has been carried out during the last 100 years in northern Pakistan. The foundations were laid by geologists of the Geological Survey of India during the period 1870–1900. More recent work, especially by M.A. Latif in the Galiat and by E.R. Gee in the Salt Range, has constrained the ages and regional correlation of these regions. Unravelling the regional stratigraphy has been difficult due to the effect of Himalayan deformation which has telescoped the entire region causing the juxtaposition of quite distinct rock sequences.

The stratigraphic evolution of the region involved three main phases. After the Pan–African metamorphic event that saw the assembly of Gondwana, the Indian subcontinent remained part of Gondwana throughout the Palaeozoic. The first phase therefore corresponds to the Gondwanian history of India. The second phase witnessed the break–up of Gondwana, the rifting of India from Africa and then Australia–Antarctica and India's subsequent northward drift. The Triassic–Lower Tertiary sequence therefore represents the history of passive margin development. The third phase documents the history of Asia–India collision, uplift and erosion of the Himalayas and sedimentation of the Upper Tertiary molasse series.

The stratigraphic column is punctuated by numerous unconformities. Latif (1970) describes seven unconformities in Hazara whilst Gee (1981) noted four in the Salt Range (Figs.6.7 & 6.8). Of these, three are of great importance regionally.

a) the Basal Permian Unconformity juxtaposes the Lower Permian Talchir Boulder Beds above Cambrian sediments, omitting the majority of the Palaeozoic era. During this period this part of Gondwana may have been subaerial. In Permo–Triassic times subsidence occurred,

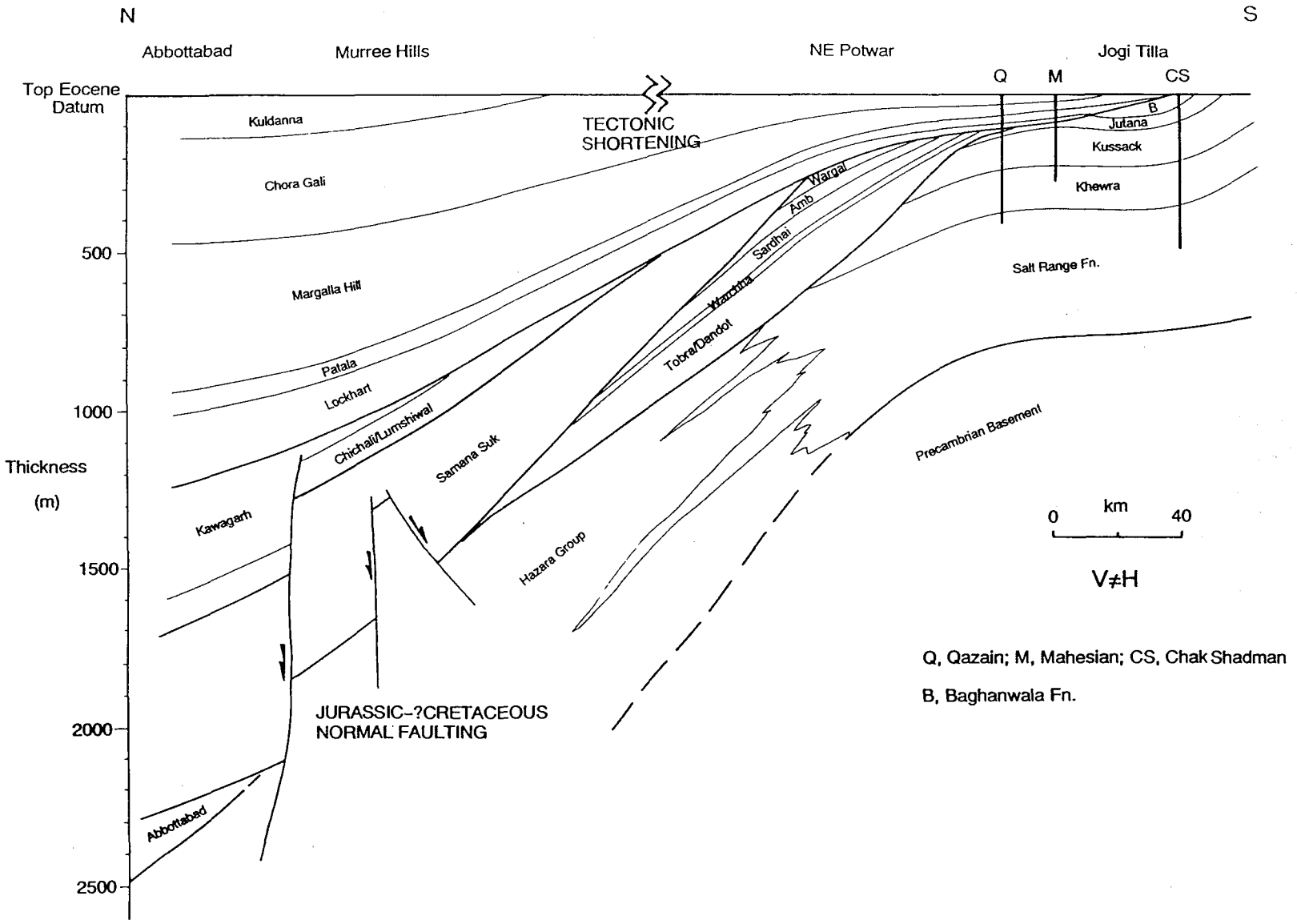


FIG.6.7. SCHEMATIC N–S STRATIGRAPHIC CORRELATION CHART AT 73 30E LONGITUDE. (incorporating unpublished O.G.D.C. data, courtesy of S.M.Naeem *pers.comm.*)

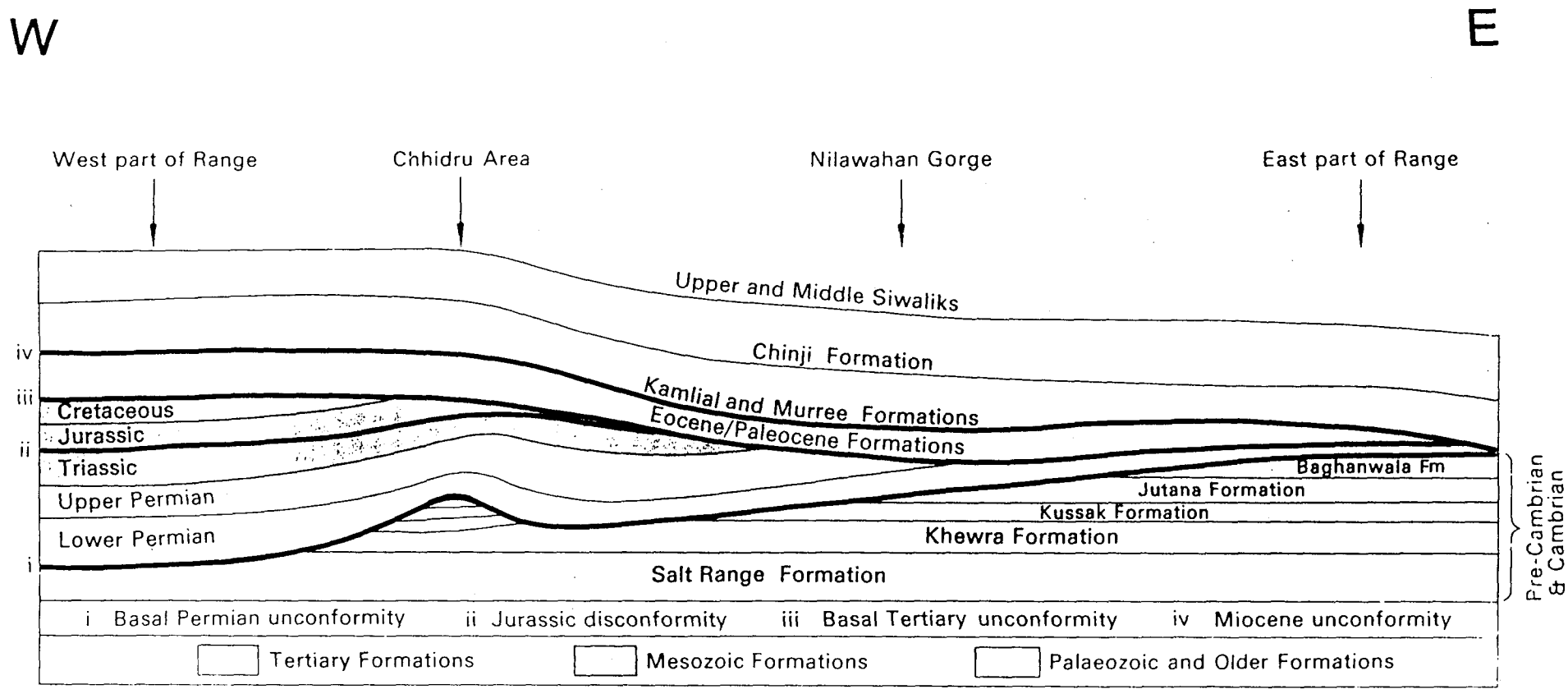


FIG.6.8. DIAGRAMMATIC E–W STRATIGRAPHIC CORRELATION CHART ACROSS THE SALT RANGES (from GEE, 1981) showing the action of the basal Permian and basal Palaeocene unconformities

perhaps related to the formation of the northern passive margin of India due to the rifting of the Lhasa block and the opening of Neotethys. Hazara lay approximately 250–500km from the Zanskar shelf at this time and therefore subsidence at this time was minor.

b) The Base Jurassic Unconformity. The Jurassic saw the creation of India's western passive margin which is associated with Triassic–Middle Jurassic rifting and Upper Jurassic spreading (156Ma) in the Somali Basin (Coster et al, 1989). This is marked by a widespread unconformity and uplift which introduced continental conditions to the Salt Range (Datta Fm.) and a condensed sequence to Hazara (Maira Fm.). A westward– and northward–thickening sedimentary wedge accumulated throughout the Jurassic and Cretaceous periods during which time unconformities appear to be of eustatic rather than tectonic origin.

c) The Base Palaeocene Unconformity. The thermal subsidence phase was interrupted by a number of regressions. The most important of these was a major uplift event at end Cretaceous–base Palaeocene times. This caused widespread regression which exposed the limits of the Zanskar shelf to subaerial erosion. Erosion during this period removed the feather–edge of the Mesozoic shelf wedge in the Potwar and is marked by lateritic and coal–bearing horizons across the entire Indo–Pak shelf. This unconformity is observed throughout Pakistan and much of India. Uplift was followed by marine transgression across the shelf and a return to carbonate deposition from Zanskar to the Salt Range. This striking regressive–transgressive cycle is thought to relate to the passage of India over the Reunion hotspot which also caused the rifting of the Seychelles–Mascarene Plateau from India and the outpouring of the Deccan Traps.

d) The Base Molasse Unconformity. The Lower Tertiary was a period of shelf sedimentation that terminated during the Eocene due to continent–continent collision. Continental conditions ensued with the deposition of approximately 5–6km of fluvial molasse sediments in the Potwar, and even greater thicknesses are found in the Sibi Trough of the Sulaiman–Kirthar foredeep. The base of this series is unconformable, the introduction of molasse sedimentation being diachronous due to the migration of the depocentre across the Indian Shield in tandem with

the advance of the Himalayan thrust front since collision in Lower Eocene times.

The following is a brief summary of the stratigraphy of the Hazara/Potwar region.

6.6.1. The Eocambrian.

The Precambrian crystalline basement is not exposed within the Hazara/Potwar area discussed here. It outcrops however towards the south on the Sarghoda "high". The oldest cover sediments that overlie this basement suite are thought to be of Eocambrian age and are represented by the Hazara Group in Hazara and the Salt Range Formation in the Potwar Plateau/Salt Range region (Fig.6.7). These units have been correlated by Latif (1973, 1984) on the basis of their relationship to the overlying Cambrian strata (Latif, 1974; Baqri, 1984) and the partial extension of evaporitic facies into Hazara (Latif, 1973). The Hazara Group, or "Slate Series" of Middlemiss (1896) and "Hazara Slate Formation" of Muhammad Ali (1962), has also been correlated with the Attock Slates of the N.W.F.P., the Dogra Slates of Kashmir and the Simla Slates of India.

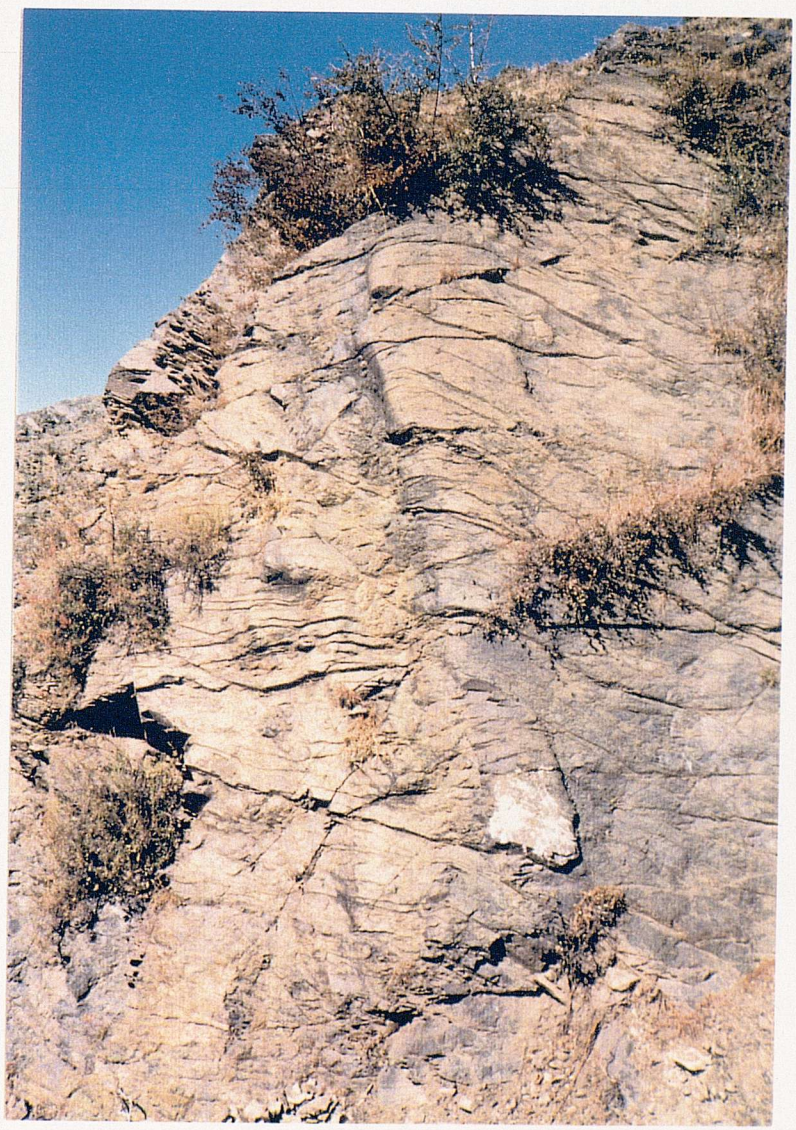
In Hazara, **the Hazara Group** comprises five formations (Latif, 1970) :

Upper Formation
Langrial Limestone
Middle Formation
Miranjani Limestone
Lower Formation

The group consists of intercalated argillaceous and arenaceous rocks including silty shales, mudstones, siltstones and greywackes (Gardezi, 1968) separated by two distinct and persistent algal limestone units. The Lower and Upper Formations are predominantly argillaceous. The Middle Formation contains an arenaceous zone more than 150m thick that forms a mappable horizon, particularly near Nathia Gali. The argillaceous material commonly displays ripple struc-



(a) Bedding–parallel foliation in the Hazara Group greywackes, N of Kalabagh Cant.



(b) Sole casts in the Hazara Group greywackes, N of Bara Gali



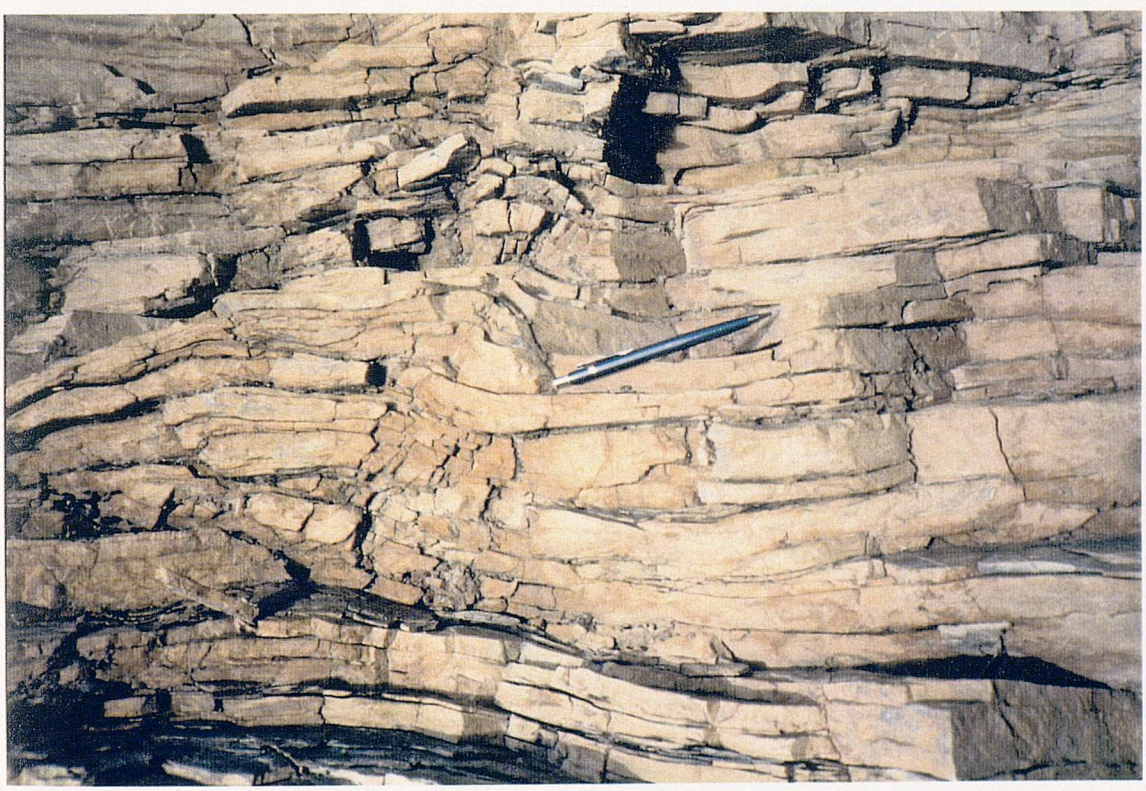
(a) Bedding–perpendicular conjugate jointing in the Hazara Group greywackes, N of Kalabagh Cant.



(b) Kinking in the Hazara Group greywackes



(a) Intense folding within the Miranjani Limestone Member of the Hazara Group close to the Hazara Detachment (MBT) at Khaner Kas



(b) Intense bedding–parallel shearing in the Miranjani Limestone Member at Khaner Kas

Mylonitic fabric in the Hazara Group close to the Hazara Detachment (MBT), just S of Muzaffarabad

tures, flute casts and dessication cracks indicating a shallow-water to intertidal depositional environment. The fine-grained beds also display a poor cleavage that is not developed in the interbedded sands. However, a strong bedding parallel fissility is present due to burial - the total thickness of the Hazara Group exceeds 3000m at Tanol (Muhammad Ali, 1962).

These units are separated by two highly distinctive algal limestones - the Langrial and Miranjani limestones. The Miranjani limestone is an algal stromatolite composed of both laterally-linked hemispheroids and spheroidal oncolites. This very distinctive rock is approximately 20-25m thick and is associated with a 10m thick gypsum band, which may be the northern extension of the Salt Range evaporitic facies (Latif, 1973). The strongly nodular Langrial limestone lies 150m from the top of the Hazara Group.

For many years the Hazara Group was regarded as unfossiliferous and therefore Precambrian in age. The discovery of Portobolella and Acrotretid within the slates (Latif, 1970), the recognition of Collenia-like algae in the limestones (Gardezi, 1968) and the Cambrian age of the overlying Abbottabad Group suggests that the Hazara Group is Eocambrian in age.

In the Potwar Plateau/Salt Range region the Eocambrian is represented by the Salt Range Formation or "Punjab Saline Series". This formation is of highly variable thickness due to halokinesis, exceeding 2100m at Dhariala (Gee, 1981), 1700m at Hayal but only 100m is recorded at Warnali (unpublished Oil & Gas Development Corporation (O.G.D.C.) data). Average thickness is approximately 400-500m. It comprises red marls and clay that alternate with gypsum/anhydrite and dolomite bands. This is overlain by thick seams of rock salt and red marls which, towards the top, includes local oil-shales and volcanics. A thin gypsum and algal stromatolite unit in the Hazara Group has been linked with this series (Latif, 1973).

The correlation of the Hazara Group with the Salt Range Formation evaporites and interbedded shales demonstrates a northward transition from salt basin to intertidal and finally shallow marine conditions (Latif, 1973). Early Cambrian subsidence was in the order of 2-3km and

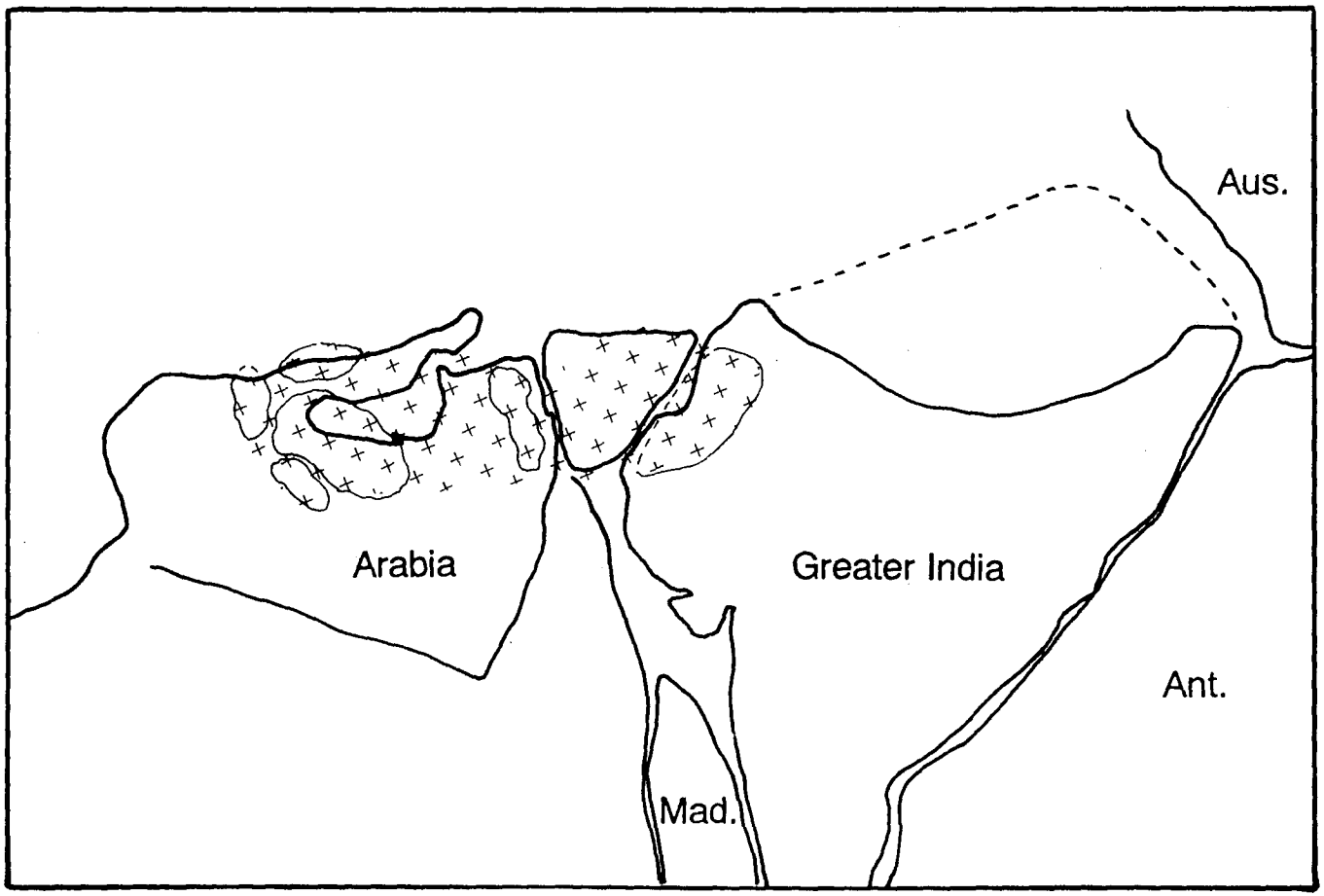


FIG.6.9. GONDWANALAND RECONSTRUCTION SHOWING THE DISTRIBUTION OF EOCAMBRIAN EVAPORITES IN PAKISTAN AND ARABIA

increased towards the north but the rate of sediment supply was sufficient to maintain a shallow marine to continental environment. The Eocambrian Salt Range Formation was deposited in a marginal evaporite basin that extended across the Helmand Block to Arabia where it is correlated with the Hormuz Formation (Shah, 1977; Zharkov, 1981) (Fig.6.9).

6.6.2. The Lower Palaeozoic.

Due to the lack of fossil evidence, the strata underlying the Mesozoic succession in Hazara has been referred to as "Below the Trias" (Waagen & Wynne, 1872), the "Infra-Trias" (Middlemiss, 1896) and more recently, the Abbottabad Formation (Marks & Muhammad Ali, 1962). This was revised to **the Abbottabad Group** (Latif, 1972) which comprises the following units :

Sirban Formation	
	Mirpur Member
Kakul Formation	Mahmdagali Member
	Sangargali Member
	Tanakki Member

This succession is 830m thick in northern Hazara (Latif, 1974) and has been correlated with the Blaini-Krol sequence in Simla, India (Fig.6.14). The Sangargali Member has been correlated with the Khewra Formation of the Salt Range (Latif, 1974; Baqri, 1984). The extension of the Sirban Formation to Kashmir has also been noted (Ashraf, Chaudhry & Qureishi, 1983) where the incomplete exposures indicate a thickness of at least 500m, which compares with approximately 570m in Hazara (Latif, 1974).

The age of the sequence has been constrained to the Cambrian on the basis of Hyolithes (Latif, 1974). In Hazara the group is unconformably overlain by rocks of Toarcian (upper Liassic) age (Davies & Gardezi, 1965). Towards the east, in Kashmir, the Sirban Formation is unconformably overlain by Palaeocene rocks (Ashraf, Chaudhry & Qureishi, 1983) whilst towards the

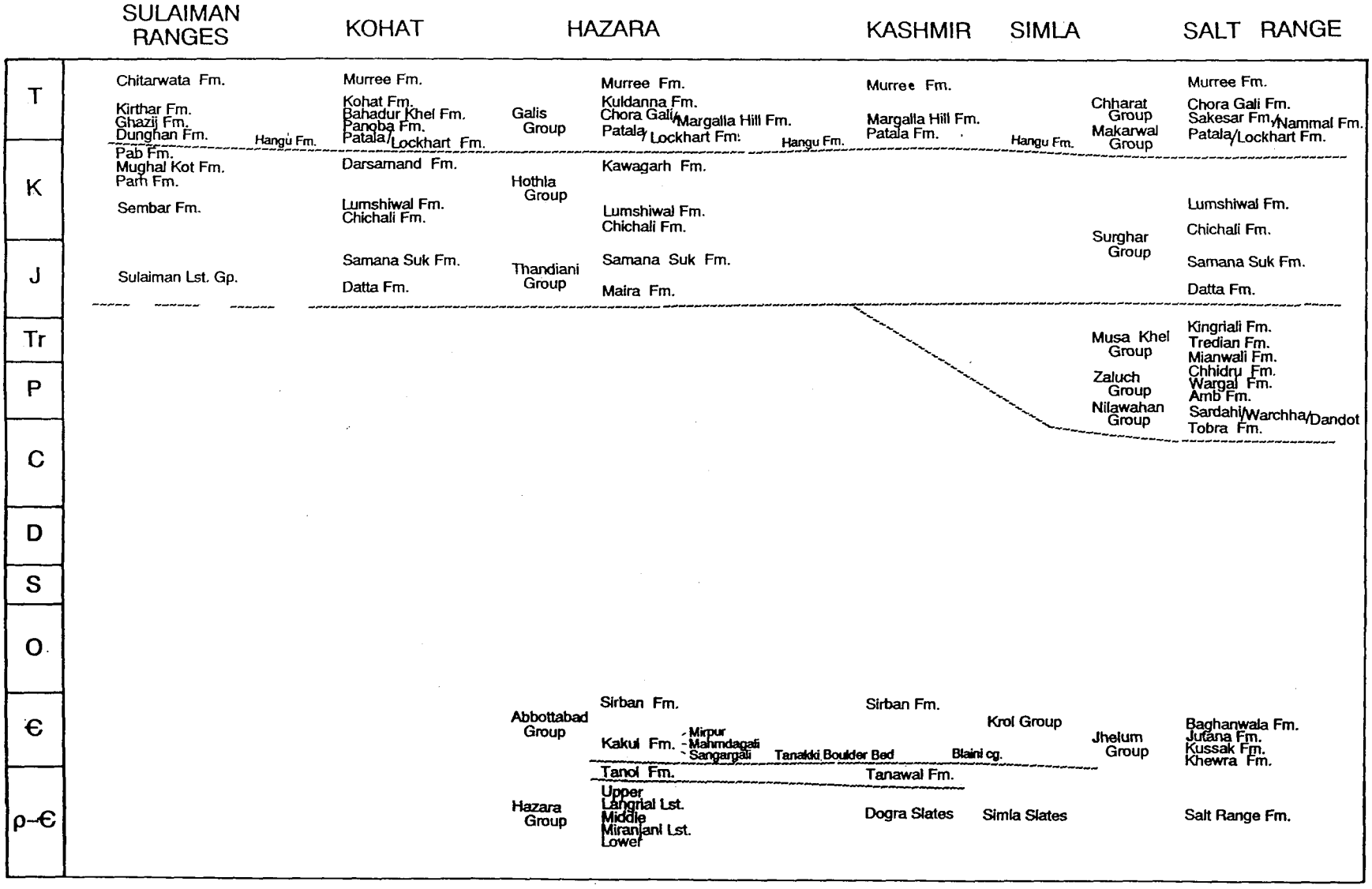


(a) Unconformity between the Hazara and Abbottabad Groups, S of Abbottabad Tanakki Cg. Hazara Group



(b) Tanakki Member, the basal conglomerate of the Abbottabad Group

Fig.6.14. STRATIGRAPHIC CORRELATION IN N. PAKISTAN



south, in the Salt Range, the equivalent Jhelum group is overlain by the Permian Talchir Boulder Bed.

The Abbottabad Group displays two shallowing–upward sedimentary cycles. The Sangar–gali Member comprises shales that grade upwards into sandstones. This is followed by the purple arenaceous dolomites of the Mahmdagali Member and then a return to clastic conditions with the arenaceous Mirpur Member. These units comprise the Kakul Formation which is approximately 250m thick in Hazara and is overlain by the 570m thick Sirban Formation, which is a thick series of unfossiliferous dolomitic limestones. This unit is uniform and widespread and contains cherty lenses (Marks & Muhammad Ali, 1961) and phosphatic horizons (Latif, 1972).

In the Salt Range, the Lower–Middle Cambrian is represented by the Jhelum Group which consists of shallow marine sands (Khewra Fm.) and shales (Kussak Fm.) that become dolomitic towards the top (Jutana Fm.) and finally evaporitic (Baghanwala Fm.) (Fig.6.14). The entire series is approximately 400m thick in the east, but is eroded by the basal Permian unconformity towards the west. The Jhelum Group comprises :

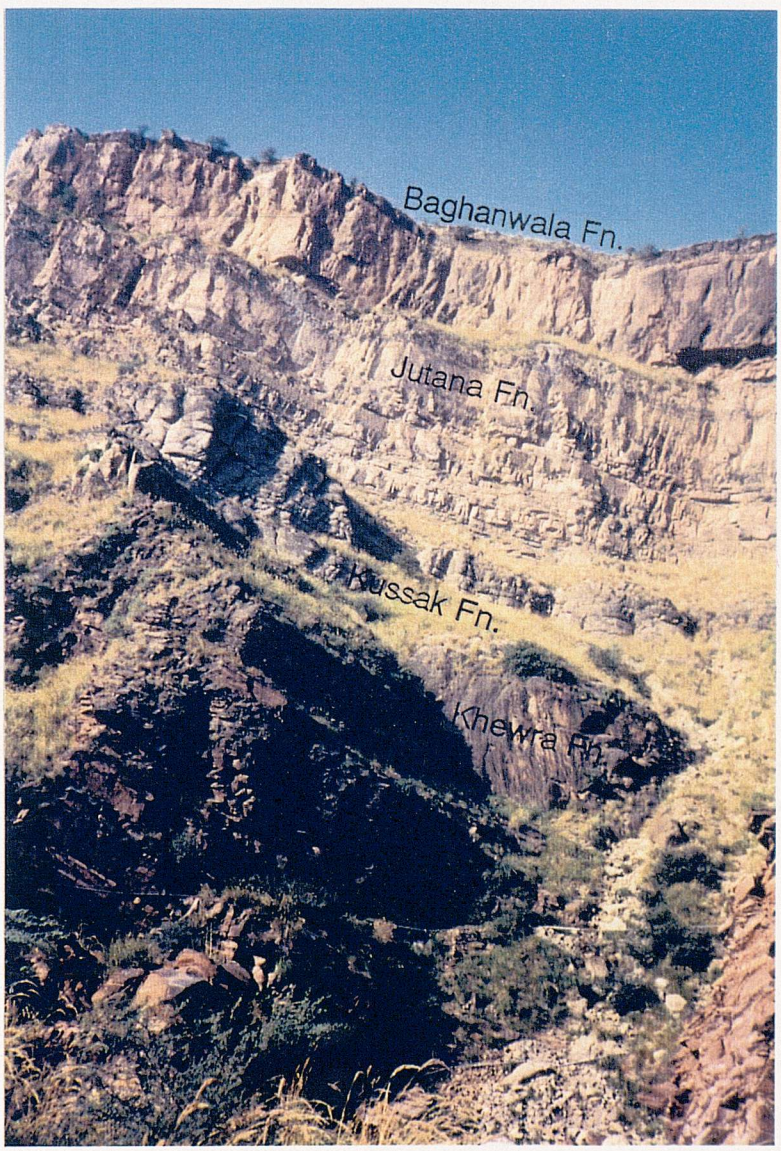
	Baghanwala Fm.
Jhelum Group	Jutana Fm.
	Kussak Fm.
	Khewra Fm.

The thick Sirban Formation of Hazara is not represented in the south. This may be due to a change in facies to the evaporitic Baghanwala Formation or the action of the base Permian unconformity in removing the uppermost Jhelum Group rocks.

The Abbottabad Group in Hazara has a narrow outcrop area which, although shortened by Himalayan thrusting, probably indicates sedimentation in a fault–bounded basin (Marks & Muhammad Ali, 1961). The group is absent in S.E. Hazara, but sedimentation is preserved in the



(a) The basal Permian unconformity N of Khewra, juxtaposing the Tobra conglomerate and the Jhelum Group



(b) The Jhelum Group, E of Khewra

Salt Range/Potwar region suggesting either compartmentalised sedimentary basins or pre-Permian uplift and erosion.

In the former case, the Jhelum Group and Abbottabad Group are seen as distinct but equivalent sequences deposited in basins separated by a non-subsiding "high". Alternatively they can be viewed as lateral equivalents within a single sedimentary basin in which subsidence halted temporarily in the north, producing the Tanakki unconformity, but then subsided throughout the Lower Palaeozoic. Then, in late Palaeozoic times uplift caused substantial erosion in the area which is now S.E. Hazara such that the Abbottabad Group was removed. This uplift may have occurred in Carboniferous times and correlate with the base Permian Talchir Boulder Bed of the Salt Range (Fig.6.10).

The Abbottabad Group unconformably overlies the Hazara Group and Tanol Formation, the erosive surface being marked by the Tanakki Boulder Bed (Middlemiss, 1896) which consists principally of poorly-rounded clasts of Hazara Group strata. In the Salt Range the equivalent unconformity is absent, the Khewra Formation overlying the Salt Range Formation with normal sedimentary contact. This reflects a sedimentary environment in which subsidence continued in the Punjab but ceased in Hazara.

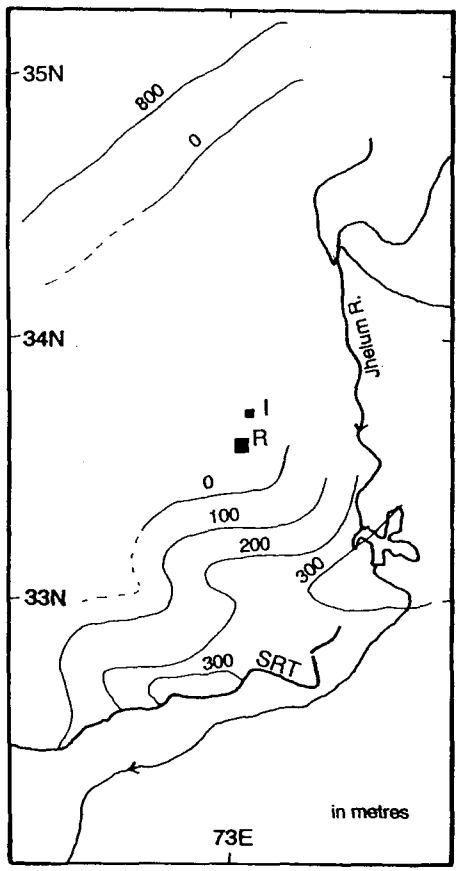
6.6.3. The Permian-Cretaceous.

The stratigraphy of Hazara is quite distinct from that of the Salt Range due to the action of unconformities, particularly the base Permian, base Jurassic and base Palaeocene unconformities; and also due to the original separation of the regions prior to Himalayan shortening. Although only 140km apart today, the Abbottabad and Salt Range regions were separated by *c.*350km prior to Himalayan collision.

The Permo-Triassic units of the western Salt Range, Potwar Plateau and Kala Chitta Ranges comprises a 1000m thick series. The basal Tobra Formation, or Talchir Boulder Bed, is a

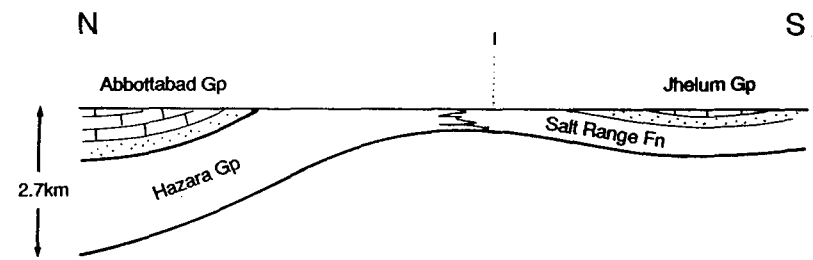
FIG.6.10. DISTRIBUTION OF LATE PRECAMBRIAN TO CAMBRIAN SEDIMENTS IN HAZARA AND THE POTWAR.

1) Late Precambrian–Cambrian Isopach Map
(with Himalayan shortening restored)

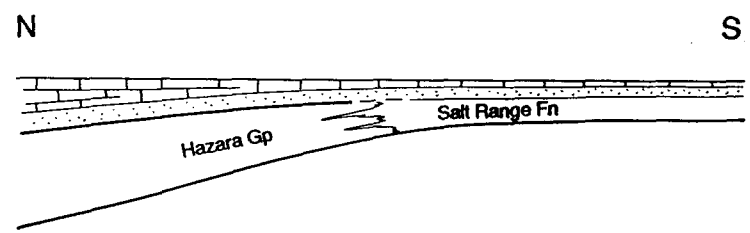


I, Islamabad; R, Rawalpindi;
SRT, Salt Range Thrust

2) Basal Permian Subcrop Section



3) Late Precambrian–Cambrian Basin Model



fluvio-glacial deposit associated with the widespread Gondwanian glaciation. This unit drapes with topographic unconformity the Jhelum Group, which had been tilted towards the southeast causing exposure of the Salt Range Formation in the west but the preservation of the Jhelum Group in the east. The same unconformity between Lower Palaeozoic strata and Dwyka glacial deposits of Gondwanaland are observed in Oman (Lee, 1990). Here, the pre-Dwyka tilting is towards the west defining a region of uplift located between Pakistan and Oman at the site of future continental rifting. Subsidence during the Permian led to the deposition of shallow marine clastics of the Nilawahan Group followed by the mainly calcareous Upper Permian Zaluch Group. Falling sea-levels brought a return to shallow marine clastic and even emergent continental clastic (Tredian Fm.) sedimentation during the Triassic. The top of the Triassic is marked by an unconformity which becomes greater towards the north (Fig.6.7). The complete Permo-Triassic sequence is therefore :

	Kingriali Fm.
Musa Khel Group	Tredian Fm.
	Mianwali Fm.
	Chhidru Fm.
Zaluch Group	Wargal Fm.
	Amb Fm.
	Sardahai Fm.
Nilawahan Group	Warchha Fm.
	Dandot Fm.
	Tobra Fm.

In Hazara the majority of the Palaeozoic and Triassic strata have been removed by the composite unconformity (Latif, 1980) at the base of the Jurassic Thandiani Group. This unconformity has been dated as Toarcian on the basis of Bouleioceras fossils in the overlying Maira

formation (Davies & Gardezi, 1965). In S.E. Hazara, Jurassic strata lie directly upon Eocambrian Hazara Group rocks, but towards the northwest the Cambrian Abbottabad Group is preserved beneath the unconformity.

The Thandiani Group comprises a basal condensed series, the Maira Formation, and a thick limestone unit, the Samana Suk Formation :

Thandiani Group	Samana Suk Formation
	Maira Formation

The Maira Formation comprises a condensed sequence of intercalated quartzite, ortho-quartzite, shales, marls and thin limestones. It has been correlated with the mainly continental Datta Formation of the Kala Chitta and Salt Ranges and the Shinawari Formation of Kohat. The Samana Suk Formation is a well-bedded, grey limestone that is dolomitic and arenaceous in its lower part. This widespread unit is capped by a laterite in Hazara (Latif, 1970) indicating a break in deposition which has not been documented in the Salt Ranges (Gee, 1981).

The succeeding Jurassic-Cretaceous series is called the Hothla Group in Hazara (Latif, 1970), which correlates with the upper part of the Surghar Group in the Salt Range (Gee, 1981). In the eastern Salt Range and eastern Potwar Plateau the Cretaceous section has been eroded and is represented by the basal Palaeocene unconformity (Fig.6.11). The Hothla Group comprises the following :

	Kawagarh Fm.
Hothla Group	Lumshiwal Fm.
	Chichali Fm.

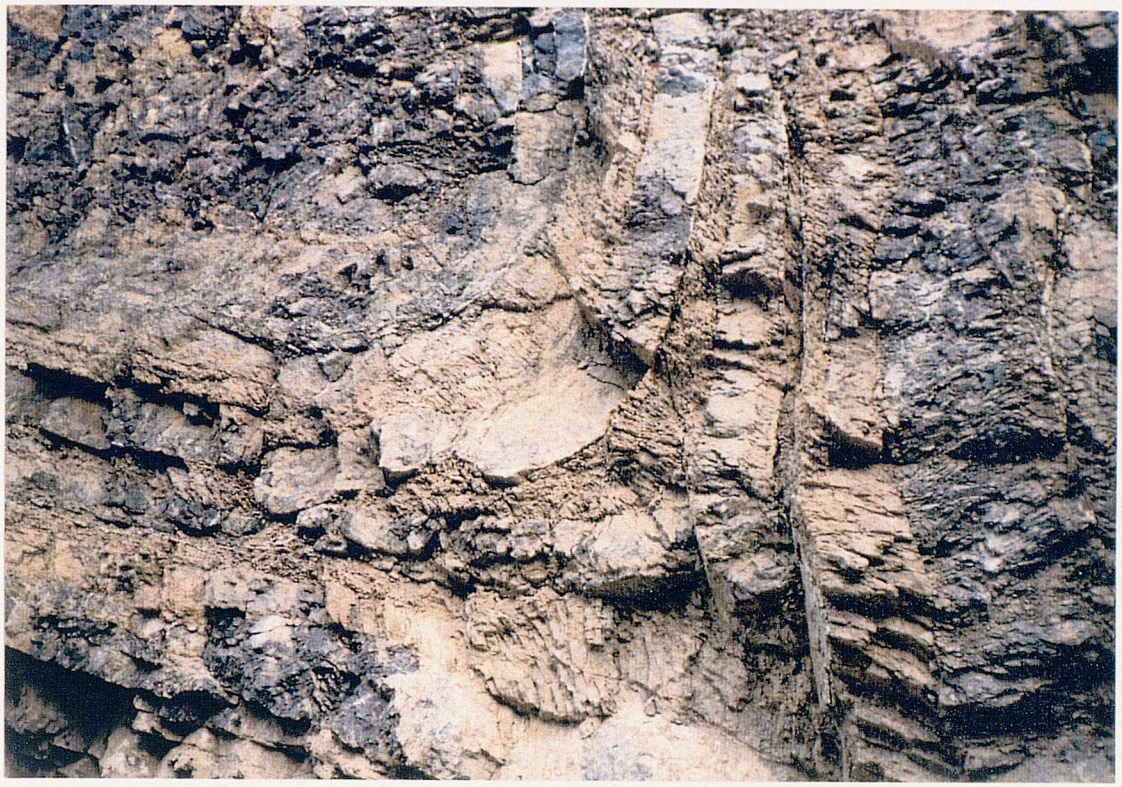
The formations have been widely correlated with the successions of Zanskar, Ladakh and Simla. The Chichali and Lumshiwal Formations are transgressive. Marks & Ahmad (1962) dated



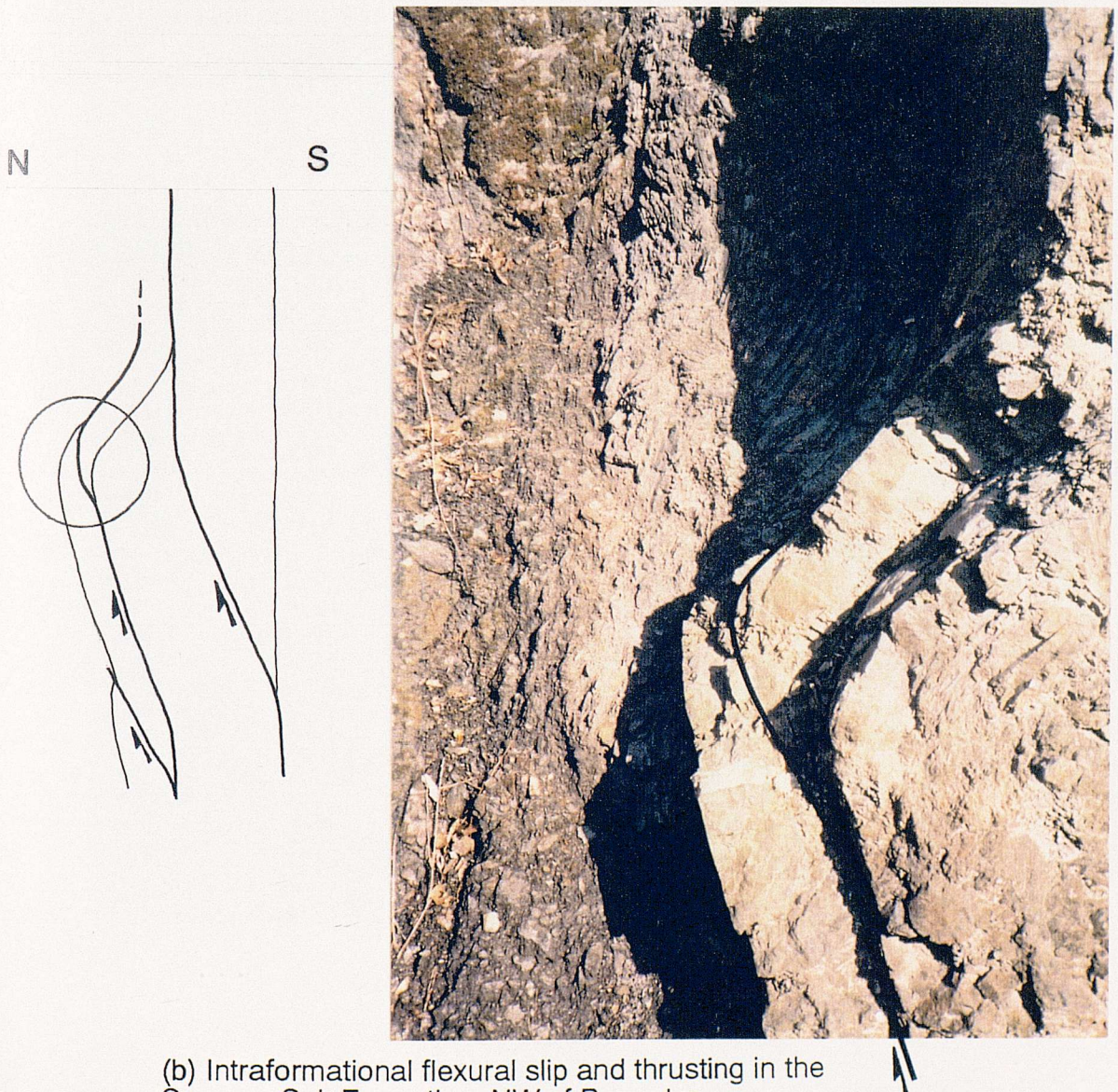
(a) The Jurassic Samana Suk Formation, SE of Changla Gali



(b) Early calcite veins offset by flexural slip during folding of the Samana Suk Formation, NW of Bagnotar



(a) Fold hinge accommodation and cleavage development in the Samana Suk Formation, NW of Bagnotar



(b) Intraformational flexural slip and thrusting in the Samana Suk Formation, NW of Bagnotar



Basal portion of the Kawagarh Formation at Kundla
(a) showing well-developed bedding



(b) showing bioturbation and pyritisation

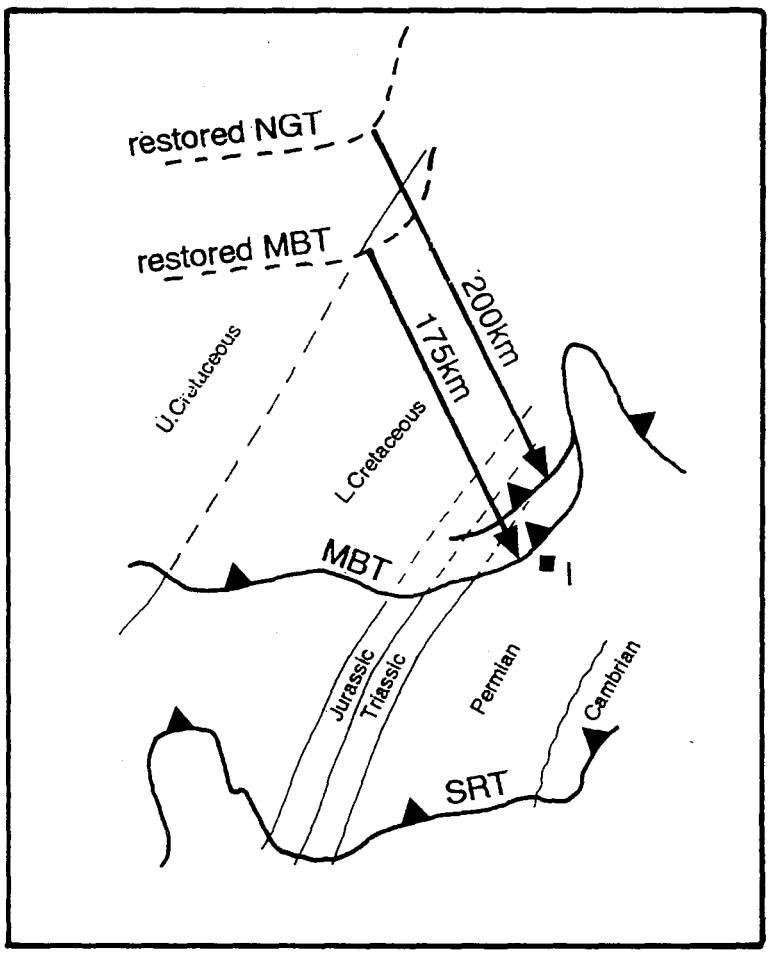


FIG.6.11. BASAL PALAEOCENE SUBCROP MAP (after restoration of MBT & NGT).



(a) The Cretaceous–Tertiary boundary between Khaira Gali and Changla Gali



(b) Marly Kawagarh Formation (locality as above)

the faunal assemblages of the Lumshiwal (or Giumal) Formation in Hazara and deduced a westerly-directed transgression - the strata younging from Valanginian/Hauterivian to the northeast of Bagnotar, to Berriasian near Nathia Gali, to Middle-Upper Albian to the southwest of Murree. The sequence also thins from 30m to 15m in this direction. After tectonic shortening of the area is restored, however, it appears that the true transgression direction may have been towards the south. This is supported by the fact that the Lumshiwal Formation in the Salt Range is fluvial rather than marine.

The Kawagarh Formation (Shah, 1977), or Chanali Formation (Latif, 1970), limestone has been dated as Upper Coniacian-Upper Campanian on the basis of diagnostic microfossils (Latif, 1962). This unit therefore was laid down during the Upper Cretaceous worldwide sea-level high. Although widespread, its distribution has been strongly attenuated by the basal Palaeocene unconformity such that it is absent in much of the southeastern Potwar and Salt Range.

In addition the formation also shows dramatic facies variations - being a thin greenish-buff, soft-weathering marl to the south of Changla Gali and a thick, white, hard-weathering limestone to the north. Such variations probably result from syn-sedimentary tectonism. The maximum reported development of the Kawagarh Formation is 90-120m (Latif, 1962) where it is a very fine-grained limestone containing pelagic foraminifera. Its basal portion is a 2-3m thick orange sandy limestone of Cenomanian age whilst its upper surface is eroded to form a laterite. However, the author has measured sections of almost 300m thickness in Ayubia. This figure may be exaggerated by tectonic thickening.

It is postulated that the Cenomanain transgression exceeded the limits of the Salt Range and drowned the Punjab Plains. However, uplift at the end of the Cretaceous prevented preservation of these deposits.

6.6.4. The Lower Tertiary.

The Palaeocene strata of both Hazara and the Potwar/Salt Range areas have been correlat-

ed (Fig.6.14). The base of the sequence is marked by the Hangu Formation which represents the most widespread unconformity in Pakistan. In the north of Hazara the Hangu Formation overlies the Upper Cretaceous Kawagarh Formation with very little stratal omission. However, late Cretaceous uplift becomes more noticeable in the south where, in the eastern Salt Range, the coaly Hangu Formation overlies Cambrian Baghanwala Formation (Fig.6.11). Erosion prior to the Palaeocene transgression is particularly marked in a chronostratigraphic diagram (Fig.6.12). At this time marine conditions were absent from the entire continental shelf since the Hangu Formation can be correlated with the Stumpata conglomerate on the Zanskar shelf (Garzanti et al, 1987). After this phase of uplift, in which almost 1000m of Permian–Cretaceous sediments were eroded (as is seen in an W–E traverse of the Salt Range), shelf sedimentation resumed until the closure of Neotethys soon after Asia–India collided.

Latif (1970) classified the Palaeocene and Lower Eocene strata as the Galis Group. In the Salt Range a different nomenclature is used, the Palaeocene being represented by the Makarwal Group and the Eocene by the Chharat Group. The Galis Group was recognised as the "Nummulitic Limestone Series" by Wynne (1874). It comprises :

	Kuldanna Fm.
	Chora Gali Fm.
Galis Group	Margalla Hill Fm.
	Patala Fm.
	Lockhart Fm.
	Hangu Fm.

The Lockhart Formation is a dark grey, nodular, bituminous limestone that weathers to a creamy–grey colour. The microfossil fauna has been dated as Lower–Middle Palaeocene. This is overlain by a buff/khaki to dark grey/black, Upper Palaeocene shale which has been correlated with the Panoba shale in Kohat. These beds also outcrop in the Salt Range where they thicken



(a) Patala Formation shales cleavage indicates overturned bedding



(b) Margalla Hill Formation nodular nummulitic limestone, W of Khaira Gali



(a) Chora Gali Formation nodular, platy-bedded, nummulitic limestone, on Ghora Gali-Lora road



(b) Kuldanna Formation shales with interbedded sands, gypsiferous marls and limestones, SE of Darya Gali

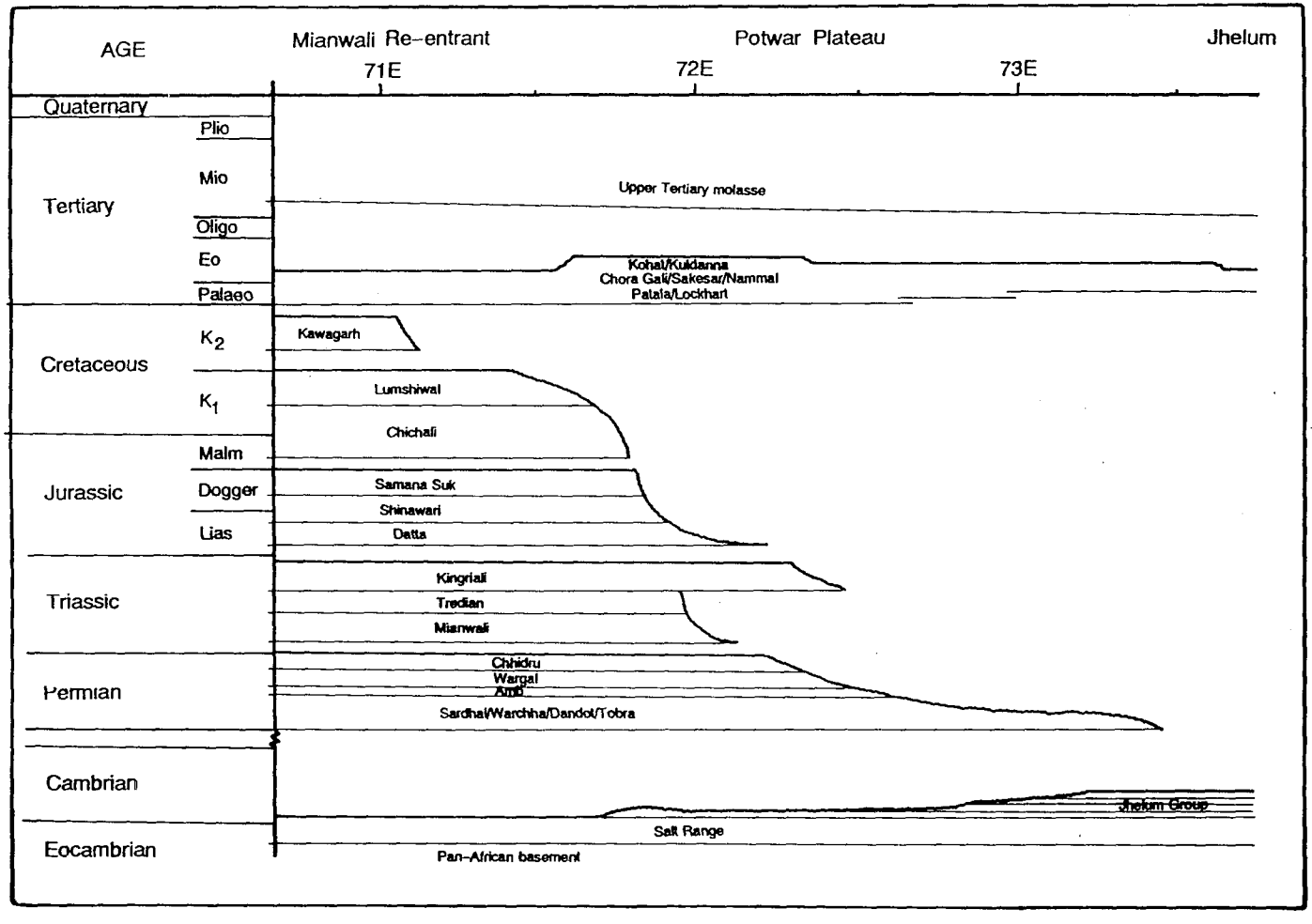


FIG.6.12. CHRONOSTRATIGRAPHIC CHART ACROSS THE POTWAR AT LATITUDE 33N.

towards the west. In the eastern Salt Range the Patala shales become coal-bearing.

The Lower Eocene sequence is highly-variable in northern Pakistan. Numerous formations have been identified from the deepwater shales of the Ghazij Formation in the Sulaiman Ranges to the shallow marine limestones of the Hazara/Potwar/Kashmir region. In between, in the Kohat region, evaporitic facies such as the Bahadur Khel salt and the Jatta gypsum are found. Two depocentres can be recognised (Latif, 1986). In the west in the Sulaiman Ranges, deepwater conditions prevailed. In the east in the Kohat/Hazara/Potwar/Kashmir region restricted lagoonal conditions were linked to the open sea towards the east. Between these basins a barrier existed, aligned with the Sarghoda basement high, where no Lower Eocene deposits occur. This represents a time of regression and can be interpreted as uplift associated with the initial collision of India and Asia at this time.

The Lower Eocene is represented in Hazara by the Margalla Hill Formation, which is a thick-bedded, nodular limestone that forms resistant bluffs and is often quarried. It is characterised by the presence of large foraminifera of early Lower Eocene age. The overlying Chora Gali Formation is a more thinly-bedded, nodular, fossiliferous limestone interbedded with pale grey/green marls. Although thin in the southeast, the Eocene strata thicken to approximately 200m in the north and west reflecting the subsidence of the Indian shelf.

The uppermost unit, the Kuldanna Formation, is transitional between marine and terrestrial conditions and marks the end of the Neotethyan realm in the western Himalayas. It comprises variegated crimson/purple shales and marls interbedded with nodular limestones and gypsiferous lenses. Blocks of serpentinite, thought to derive from the suture zone, are also reported (R. Graham & R.Beck, pers. comm.).

In the Kala Chitta and Kohat regions the top of the Eocene series is marked by the Kohat Formation which comprises white limestone containing very large Nummulites fossils. This unit is absent in the east reflecting southwestward progradation of the molasse basin and the survival

of marine shelf conditions later in the southwest than in the northeast. The uppermost Neotethyan sediments were exposed and re-worked prior to the influx of the terrigenous molasse deposits. The base of the molasse is therefore marked by the Fatehjang Member which consists of detrital Nummulites fossils enclosed in a sandy matrix. This diachronous bed represents an erosive period as the Himalayan forebulge migrated across Hazara towards the foreland.

6.6.5. The Upper Tertiary.

The Upper Tertiary sequence comprises the fluvial molasse series which is divided into the Rawalpindi and Siwalik Groups. Recently, a marine tidal-flat sequence has been identified in the extreme northern exposures in the core of the Hazara-Kashmir Syntaxis (the Balakot Formation of Bossart & Ottiger (1989)). Molasse sediments are deposited in the flexural basin behind a migrating uplifted forebulge. They are therefore unconformable and transgressive, the depocentre migrating in tandem with the forebulge and the deformation front. In some foredeeps marine conditions persist long after collision, for instance the Alpine molasse basin of Austria (Nachtmann & Wagner, 1987). Under these conditions the classic transition from deepwater flysch facies to shallow marine or fluvial molasse facies sedimentation is commonly observed. In Pakistan, the preservation of the molasse basin is very incomplete. Until recently only Miocene and younger fluvial molasse had been documented. Bossart & Ottiger (1989) have reported Palaeocene age tidal marine deposits at the base of the molasse series. This suggests that the common assumption of non-deposition during the Oligocene is false and that these sediments were deposited, within what is now the High and Lesser Himalayas, but have since been uplifted and eroded.

A total of approximately 5500m of molasse sediment has been deposited in the Potwar basin (Lillie et al, 1987) and similar thicknesses are estimated in the Gangetic foredeep, but this figure increases dramatically in the N-S trending Sulaiman and Kirthar foredeeps.

a) The Rawalpindi Group.



(a) Shale–dominated Murree Formation cycle channel sand with erosive base overbank sequence of shales and thin sands



(b) Fatehjang Member comprising 100% detrital nummulites and marking the onset of molasse deposition at Kuldanna

The Lower Miocene Rawalpindi Group lies unconformably upon the Eocene substratum throughout the area under discussion, although further to the south it onlaps onto Cambrian strata. It is exposed in the Galiat, northern Potwar and in the strongly uplifted structures of the southern Potwar and the Salt Range. It varies in thickness from 4–5km in the north to less than 500m in the eastern Salt Range and it is not thought to occur in the Warnali well (unpublished O.G.D.C. data) to the south of the Salt Range. Thickness data are viewed cautiously due to the lithostratigraphic basis of molasse nomenclature in the Himalayas.

Sub–division of the molasse series is arbitrarily done on the grounds of sandstone : shale ratio. Clearly, in a prograding fluvial series this criterion varies not only temporally but also spatially as one moves away from the main channel system towards overbank areas. Lithostratig–raphy, although a useful local field tool, is not helpful in correlating the molasse on a regional scale. For this, a chronostratigraphical method such palaeomagnnetic reversal stratigraphy is more useful (Raynolds et al, 1980).

The Group comprises the following :

Rawalpindi Group	Kamlial Formation
	Murree Formation

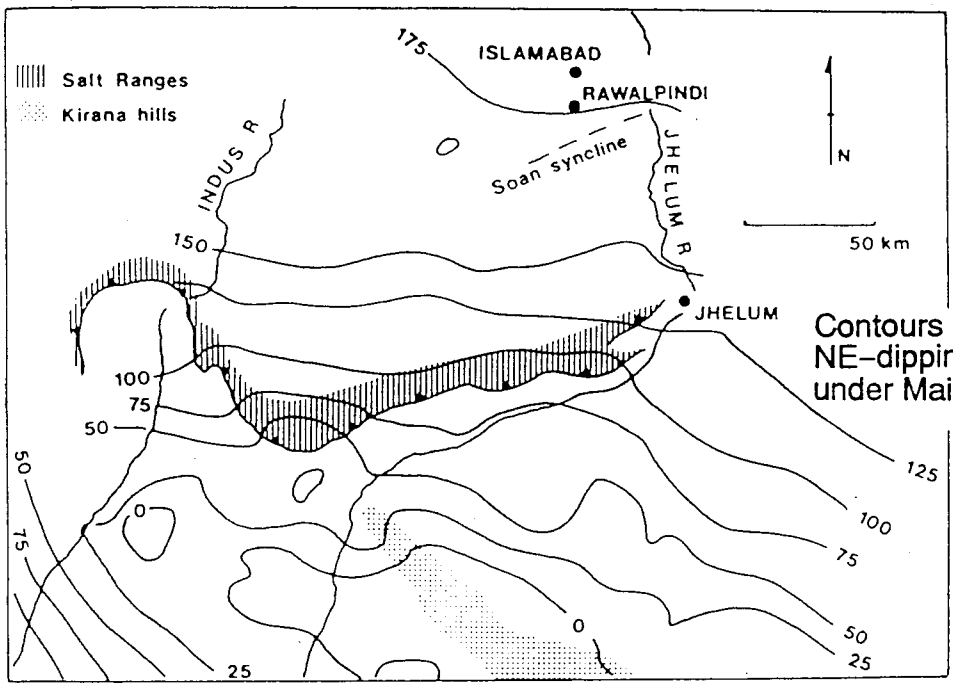
The Murree Formation consists of alternating greyish–green sandstones and red–purple silty shales deposited in a fluvial environment. Petrographic analysis suggests a Himalayan source for the Murree Formation, although early workers (Wadia, 1932) suggested a southerly provenance area on the Indian Shield. The sandstones are composed principally of quartz and carbonate fragments but also contain volcanic fragments with an island–arc association. This suggests that the provenance area was the Kohistan–Ladakh region and the uplifting Himalayas rather than the Indian Shield.

Gravity data suggests that the molasse basin trends NW–SE parallel to the main Himalayan arc and deepens towards the mountain front (Fig.6.13), and this is bourne out by reflection seismic data. The oldest part of the Murree Formation therefore was deposited in the northeast. It has been suggested that this part of the formation is marine (Bossart & Ottiger, 1989), and accummulated in a tidal–flat environment. As orogeny continued and marine conditions were eliminated the Murree Formation became fluvial. Hence, in the northeast, the carbonate shelf sequence is replaced by terrigenous clastics which eventually grade into fluvial molasse. This transitional sequence was recognised by Wadia (1932). In the southwest there was a period of non–deposition after the cessation of shelf sedimentation and the overlying fluvial molasse is unconformable.

The marine part of the Murree Formation consists of numerous fining–upward cycles, each being 1–20m thick. Six facies have been recognised within each cycle ranging from basal conglomerate to coarse cross–bedded sandstones which grade upwards into finer–grained rippled or laminated sandstones and finally siltstones (Bossart & Ottiger, 1989). The conglomerates which have strongly erosional bases, can be up to 2m thick. They may consist of siltstone flakes derived from the underlying cycle or more rounded silt or micritic limestone pebbles derived from previously deposited cycles. The channel units consist of coarse–grained lithic greywackes that may reach 10m thickness. They display trough cross–stratification on a 50cm scale. Bioturbation is common. Towards the top of the sandy units the grain size diminishes and ripple cross–stratification is gradually replaced by interbedded sand/mud laminations indicative of a tidal environment. Bossart & Ottiger (1989) describe this feature as "tidal bedding" (Reineck & Singh, 1967). The overlying siltstone facies retains no sedimentary layering.

The fluvial molasse shows a similar character but the sedimentary structures are assymmetric due to the monopolar current direction, in contrast to the bipolar tidal–flat environment.

Bossart & Ottiger (1989) mapped a 15000m thick series of Murree Formation molasse. This seems exaggerated and suggests that tectonic repetition was overlooked. It is unlikely that



Bouguer gravity map of northern Pakistan (after Farah *et al.* 1977). The units of the contours are milligal (1 mGal = 10^{-3} cm s^{-2}).

FIG.6.13. BOUGUER GRAVITY MAP OF THE NORTHERN PUNJAB (from Coward et al, 1988).

the thickness of the Murree Formation exceeded 6000m.

b) The Siwalik Group.

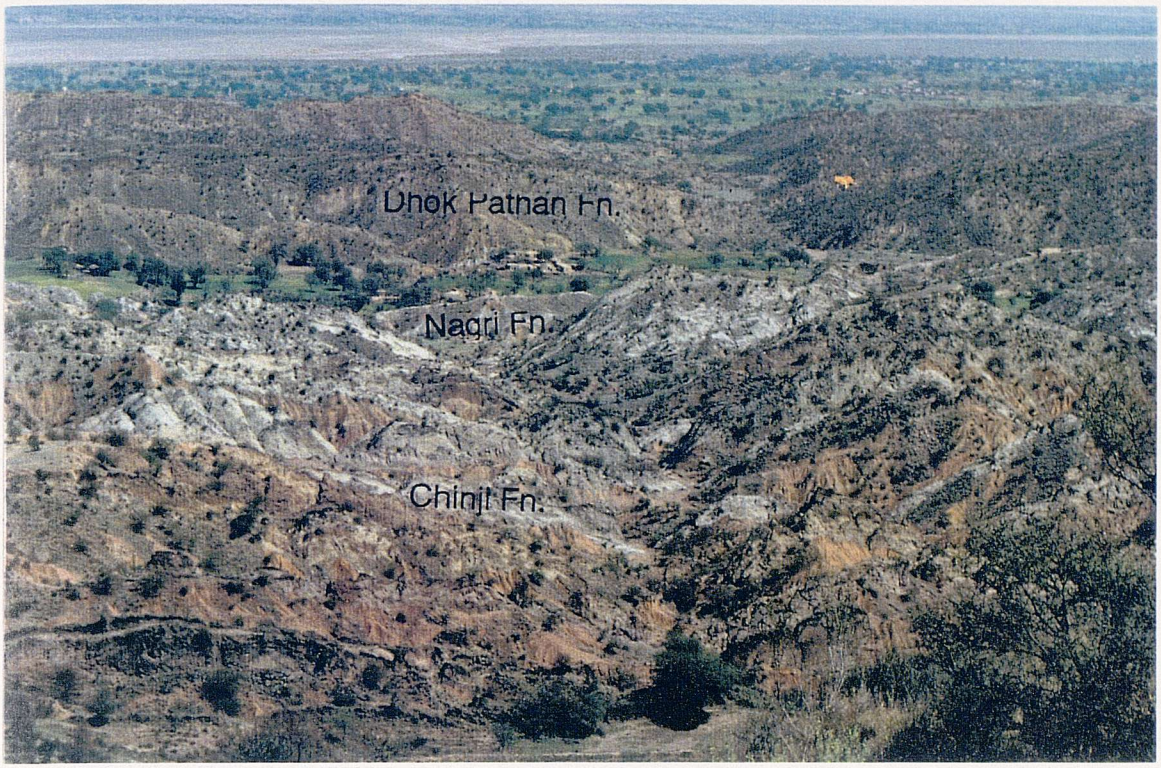
The Siwalik Group is in gradational contact with the underlying Rawalpindi Group molasse - the boundary between them being arbitrarily defined as the point at which the sandstone/shale ratio falls below 50%, marking the top of the Kamlial Formation and the base of the Chinji Formation. The Siwaliks comprise the following formations :

	Soan Formation
Siwalik Group	Dhok Pathan Formation
	Nagri Formation
	Chinji Formation

These formations prograded southwards and southwestwards as the Himalayan deformation front advanced.

The Siwaliks of the Potwar are lithologically less mature than the Murree Formation strata indicating a closer proximity to the source area. The proportion of quartz has fallen from 30% to 25% whilst the carbonate fragments have increased from 25% to 30% compared to the Rawalpindi Group. Magnesium, calcium and iron oxides have all increased in the Siwaliks indicating the unroofing of the internal zones of the Himalayas.

As one rises through the Siwalik section the intensity of colour contrast, which is related to post-depositional oxidation, decreases. The Chinji Formation is characterised by thick red/scarlet overbank shales that are interbedded with subordinate grey channel sandstones. Chinji channels are usually 5-10m deep and trough-bedded dunes are less than 50cm high. The shales are typified by shrinkage, dessication and synaerisis cracks and ochreous/orange palaeosols with



(a) View of the Siwalik Group from Chambal ridge, eastern Salt Range.



(b) Conglomerate unit in the Soan Formation demonstrating the introduction of more proximal molasse conditions

carbonate-rich nodules below. The Dhok Pathan Formation, however, although having a similar sandstone/shale ratio is characterised by dull pink overbank deposits and brownish-buff sands.

The intervening Nagri Formation consists of three to four major channel units separated by thin overbank facies. The base of the Nagrigri Formation is marked by a reversal in current direction from E-NE flowing in Chinji times to SW-flowing at the base of the Nagri. This direction then reverses again in the upper Nagri. Compared with the Chinji Formation, the Nagri Formation is typified by thicker multi-storied sandstone bodies that are 30-50m thick and reach 70m thick in the type area. The individual sandstone stories are also thicker, perhaps 10-20m thick, indicating deeper river channels; and the interconnected ratios of the sandstones are increased, as is river sinuosity.

The topmost unit, the Soan Formation, is dominated by much coarser sediments including major conglomeratic units that prograde towards the foreland with time (Raynolds, 1981).

The above description is plagued by the fact that the Siwaliks are the result of a complex fluvial system. Therefore, generalisations relating to sandstone/shale ratio depend also upon the proximity to the main channel system. A thick, rapidly-accummulating channel sand body may correlate along strike with a thin overbank shale sequence or even a palaeosol. Regionally correlative litholgical horizons are rare, although some e sands have been mapped for 20-30km across the Potwar (Behrensmeyer & Tauxe, 1982). Palaeomagnetic reversal stratigraphy has become the most reliable correlation tool since it permits a chronostratigraphic framework to be established in a region where facies vary rapidly spatially (Johnson, Raynolds & Burbank, 1986; Burbank, Raynolds & Johnson, 1986).

The use of chronstratigraphy allows us to estimate time-averaged sedimentation rates for the Siwaliks. During the Middle Miocene, at the time of Chinji Formation deposition, sedimentation rates equalled approximately 0.13mm/yr. Data from near the Nagri-Dhok Pathan Formation boundary yield an average sedimentation rate of 0.52mm/yr (Behrensmeyer & Tauxe, 1982).

With the increasing proximity of the mountain front the Siwalik sedimentation rate increases; a feature that was also noted in the Murree Formation by Bossart & Ottiger (1989).

CHAPTER 7.

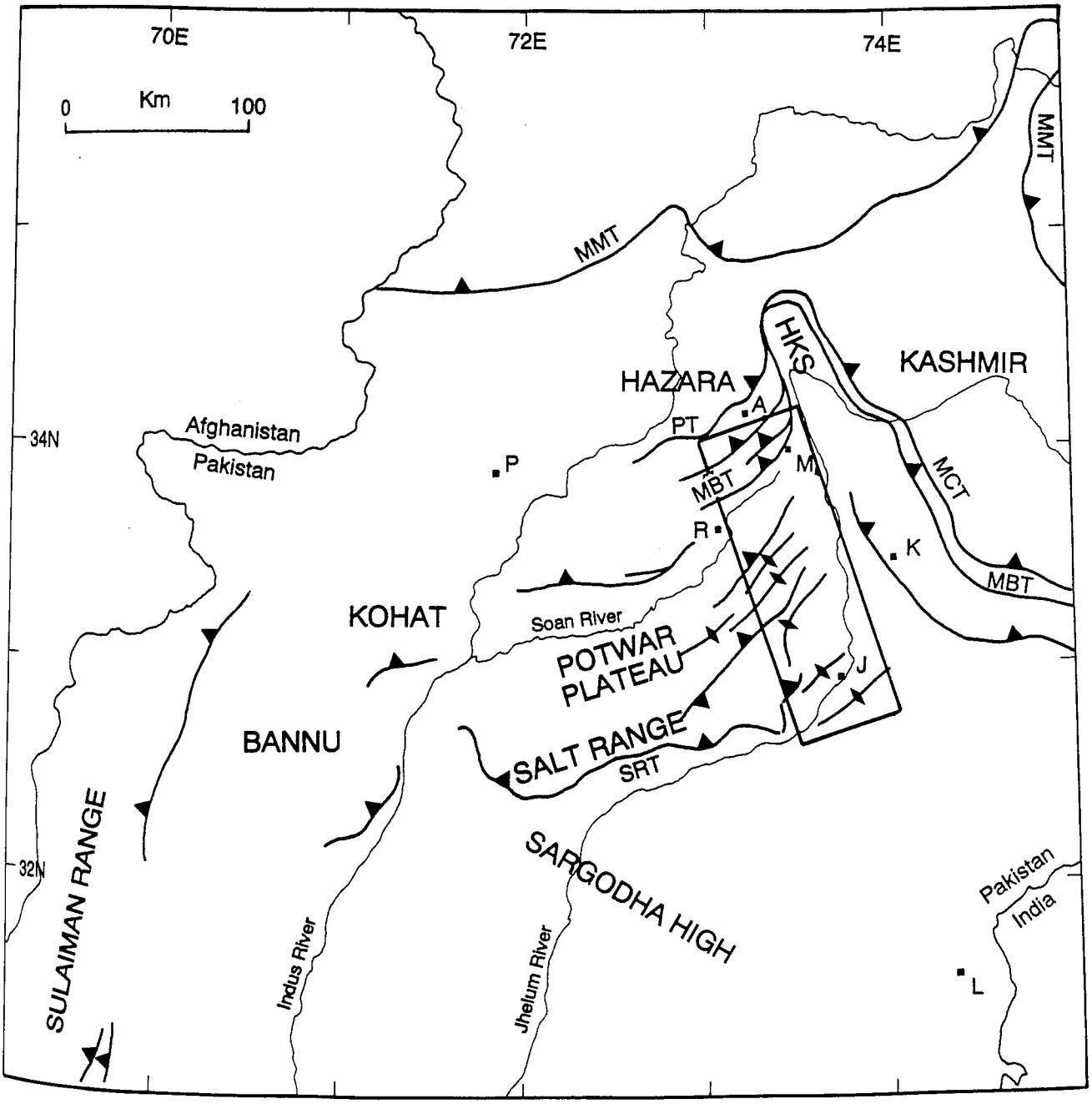
FIELD DATA FROM THE PAKISTANI HIMALAYAS.

The aim of this chapter is to present the field and seismic data from S.E. Hazara and the eastern Potwar Plateau (Fig.7.1). This region lies within the external zone of the Himalayan thrust wedge. It comprises the deformed sedimentary cover of the Indian shield that has been decoupled from its Precambrian crystalline basement. Deformation has occurred in the last 15Ma during continuing foreland basin subsidence.

The structure of the region reflects the south–southeastward propagation of the Himalayan thrust front during the last 15Ma. In the middle Miocene the Himalayan deformation front lay in the hangingwall of the MBT. Major displacement then occurred on the MBT itself, accompanied by duplexing in the footwall of the MBT, during the upper Miocene. In the early Pliocene the basal detachment of the thrust wedge encountered the Salt Range Formation evaporites and the deformation front propagated rapidly to the Salt Range thrust (SRT). Continued deformation occurred in the early Pliocene by out–of–sequence thrusting in the Potwar Plateau producing the Riwat thrust and the Soan syncline. In the upper Pliocene passive–roof duplexing between detachments at the basement–cover contact and a high fluid pressure horizon within the molasse produced intense deformation and uplift in the NPDZ. Back–folding above the roof backthrust of this duplex is observed in the NPDZ and the MBT zone and is responsible for the generally steep dip of bedding at the surface in these zones. Pleistocene compression is manifest in the South Potwar where halokinesis initiated pop–up formation and fault–propagation folding. Recent SSE–vergent deformation has occurred at the SRT where Recent river gravels have been folded.

The structural data are described in three regions from north to south, 1) the *Galiat* (Section 7.1), 2) the *North Potwar Deformation Zone (N.P.D.Z.)* (Section 7.2) and 3) the *South Potwar* (Section 7.3).

Fig.7.1. LOCALITY MAP OF THE HAZARA–KASHMIR SYNTAXIS



A, Abbottabad; J, Jhelum; K, Kotli; L, Lahore; M, Murree; P, Peshawar; R, Rawalpindi;

HKS, Hazara–Kashmir Syntaxis; MMT, Main Mantle Thrust; MCT, Main Central Thrust; PT, Panjal Thrust; MBT, Main Boundary Thrust; SRT, Salt Range Thrust.

Boxed region defines study area

View of the Salt Range Thrust carrying Khewra Formation
over monoclinally-folded Siwalik sediments



(a) SSE–dipping Siwalik sediments in front of the SRT, eastern Salt Range.

7.1. STRUCTURAL DATA FROM THE GALIAT.

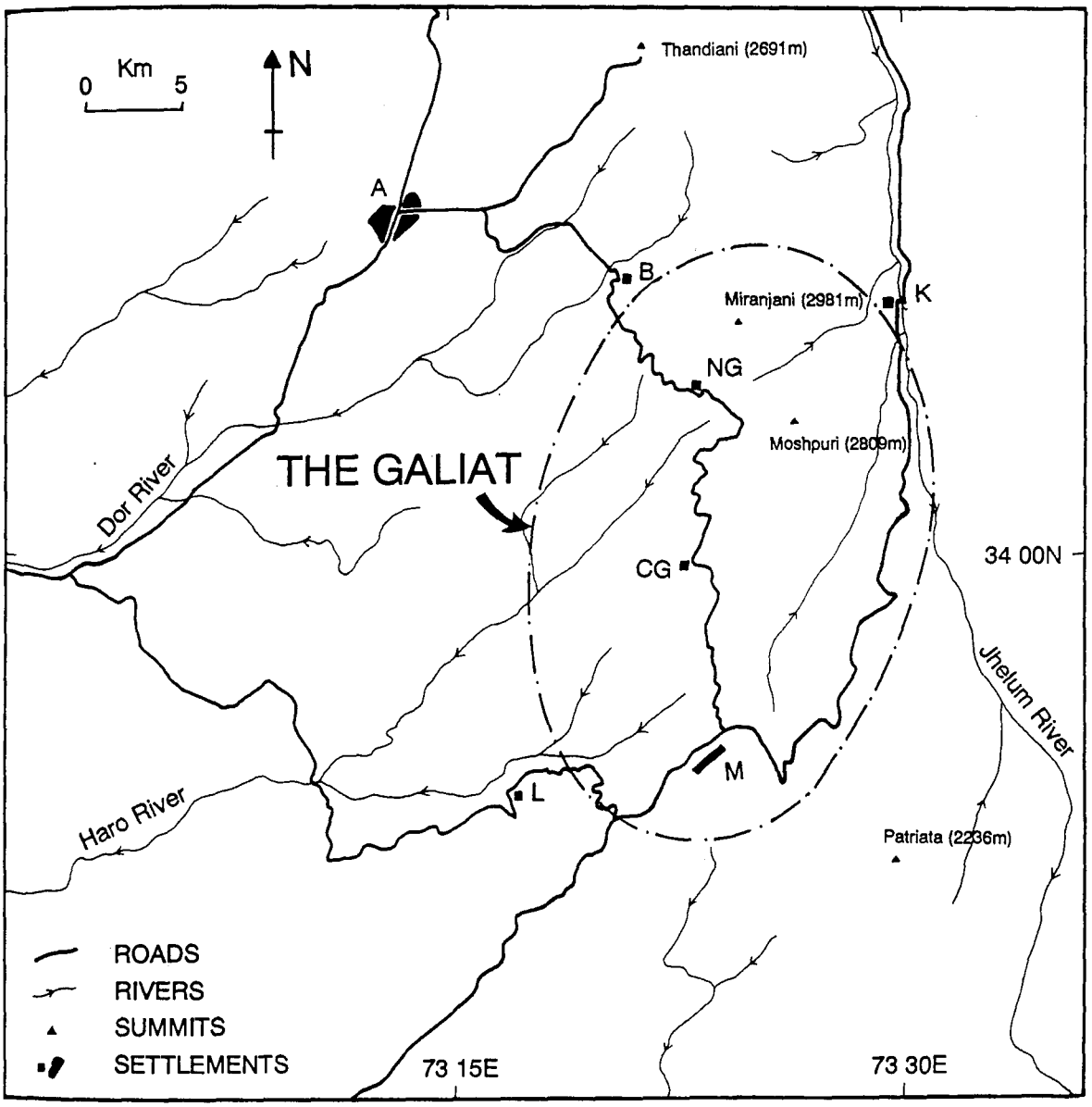
The Potwar Plateau is bounded to the north by the Himalayan foothills which are locally called the Kalachitta, Margalla and Murree Hills. In the west, the Kalachitta Ranges (1500–2000m) form a discrete range that pass northwards into the Peshawar Basin. Towards the east the foothills become topographically higher, elevations typically exceeding 2000m in the Murree Hills. This elevated region extends to the north and is known as the Galiat.

The Galiat lies within the Lesser Himalayan zone of Gansser (1964). It is bounded to the east by the Jhelum River but towards the west its boundary is vaguely defined due to the gradual decrease in altitude. Politically, it straddles the boundary between the Hazara District of the N.W.F.P. and the Rawalpindi District of the Punjab.

The Galiat is essentially defined on grounds of altitude. This is most clearly illustrated by an analysis of the drainage pattern (Fig.7.2). Two major antecedent rivers drain the region – the Indus and Jhelum Rivers – and they flow in a generally southward direction. The minor rivers, however, are directly constrained by the geological strike and the major uplift produced during the formation of the Hazara–Kashmir Syntaxis. The N–S trending watershed that separates the Indus and Jhelum catchment areas is a recent feature of Pleistocene age. It is suggested that the uplift is an expression of the blind northwestern continuation of the Kotli Thrust and therefore the frontal expression of the SW–vergent Kashmiri thrust wedge.

This recent uplift is reflected in the altitude of the region, which includes peaks such as Miranjani (2981m) and Moshpuri (2809m), and the deeply–incised drainage system. As a result there is up to 2000m of vertical exposure in the Galiat. The structure of the region has remained poorly understood, in spite of numerous stratigraphic studies, due to the often thick cover of vegetation.

Fig.7.2 TOPOGRAPHY, DRAINAGE AND COMMUNICATIONS IN THE GALIAT



A, Abbottabad; B, Bagnotar; NG, Nathia Gali; CG, Changla Gali; M, Murree; K, Kohala; L, Lora.

7.1.1. Murree

Murree lies 40km to the northeast of Islamabad at an altitude of 2200m. Once a British hill station, Murree remains an important military and administrative centre. It is a popular tourist centre, being readily accessible from the cities of Rawalpindi and Islamabad and providing relief from the heat of the Punjab plains in summer.

Geologically Murree is important because it lies on the boundary between the calcareous Lesser Himalayan zone and the molassic Sub-Himalayan zone (Gansser, 1964) or, more precisely, the Nummulitic and Upper Tertiary zones of Middlemiss (1896). Elsewhere in the Himalayas this boundary is marked by the Main Boundary Thrust (MBT). The MBT is typically an emergent forethrust that carries the Lesser Himalayas over the molasse basin. In N. Pakistan the Murree Thrust is correlated with the MBT despite the fact that, due to a backfolding phase, the Murree Thrust is rarely emergent. In the Margalla Hills the "Murree Thrust" or "MBT" remains blind and is represented at the surface by the exhumed core of an anticlinal backfold. Lateral variation of the MBT mountain front has been discussed by Izatt (1990).

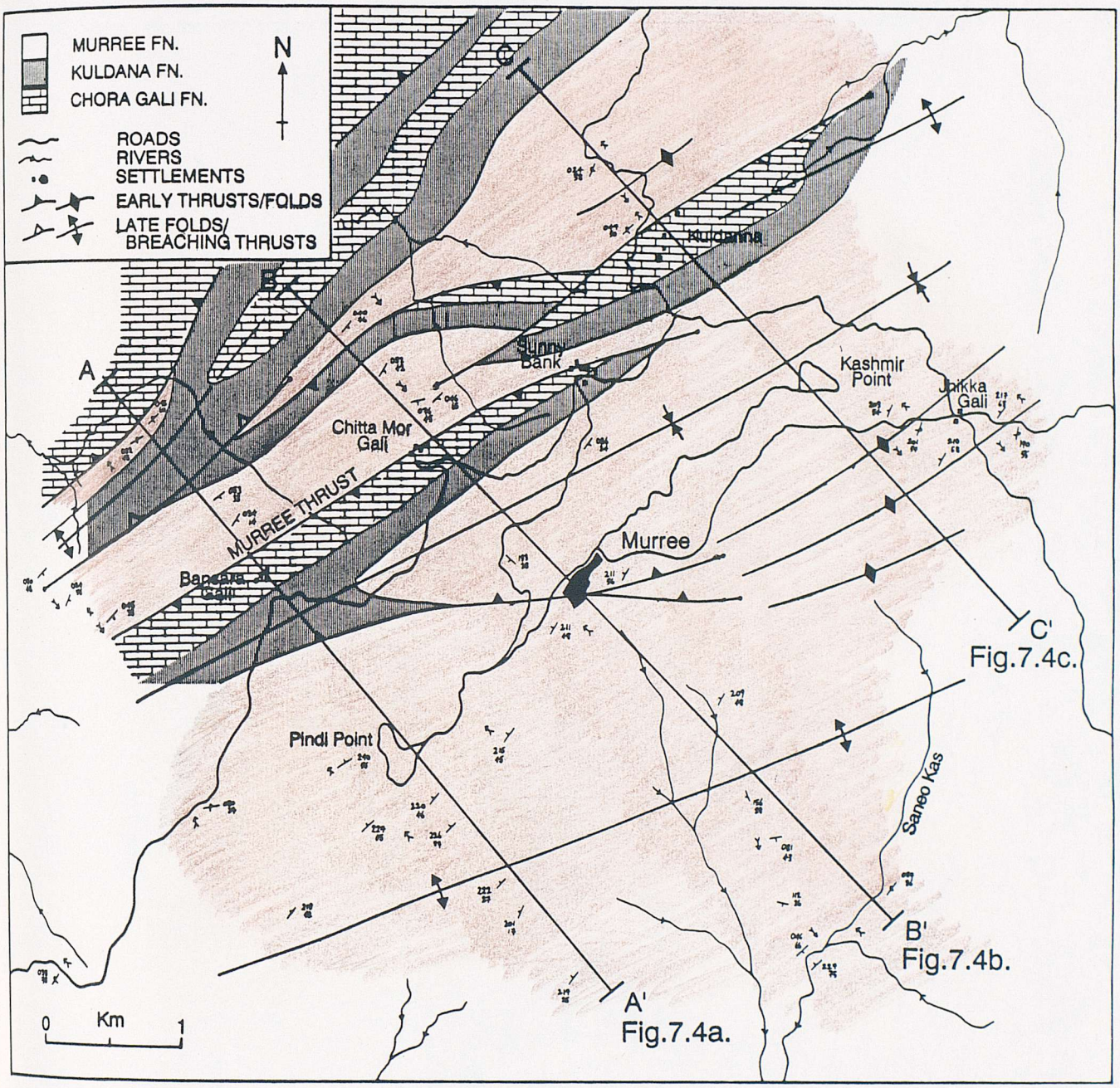
The structure of the Murree area is illustrated by a geological map (Fig.7.3) and three parallel cross-sections (Figs.7.4a,b&c). The mapped geology is superficially similar to that of Latif (1970)(Figs.7.5 & 7.6), but has been interpreted quite differently. The structure of the Murree area is typical of that observed in the rest of the Galiat. **Two deformation phases are distinguished :-**

1) Early SSE-vergent thrusting accompanied by tipfolding and

2) Late NNW-vergent backfolding and breach faulting.

The basic structure is illustrated by a section through Bansara Gali and Pindi Point (Figure 7.4a). The topographic high of Pindi Point itself is the core of a major anticlinal backfold in which a 2.5km thick section of Murree Formation molasse is exposed. The bedding youngs consistently towards the north but dips fan from 20-25°NW through vertical to 50-60°SE as one crosses the backfold axis. Tectonic thickening may have occurred. Near Bansara Gali one

Fig.7.3 GEOLOGICAL MAP OF MURREE SHOWING THE GEOMETRY OF THE MURREE THRUST



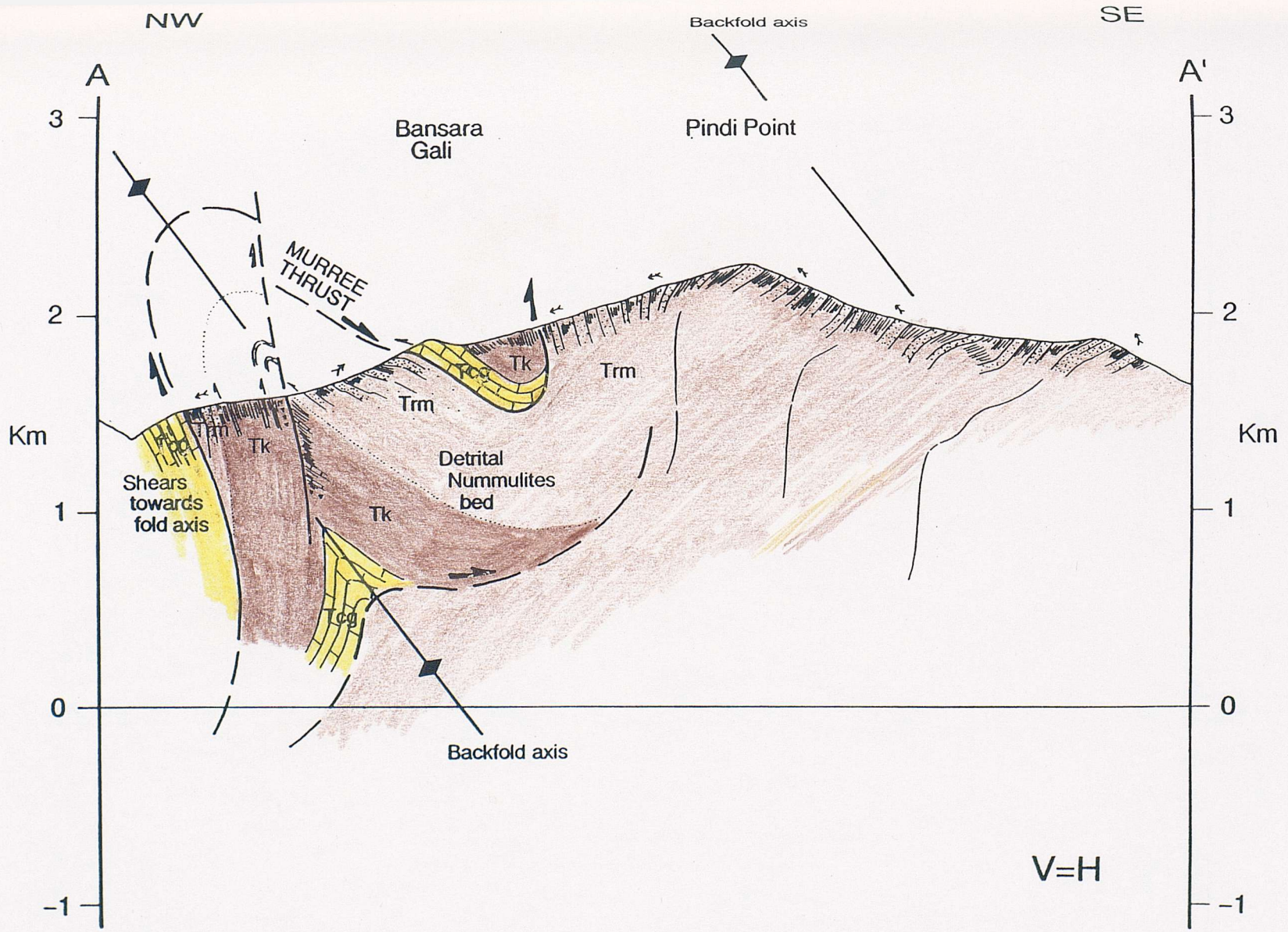


Fig.7.4a. SECTION ACROSS THE MURREE THRUST AT BANSARA GALI/PINDI POINT.

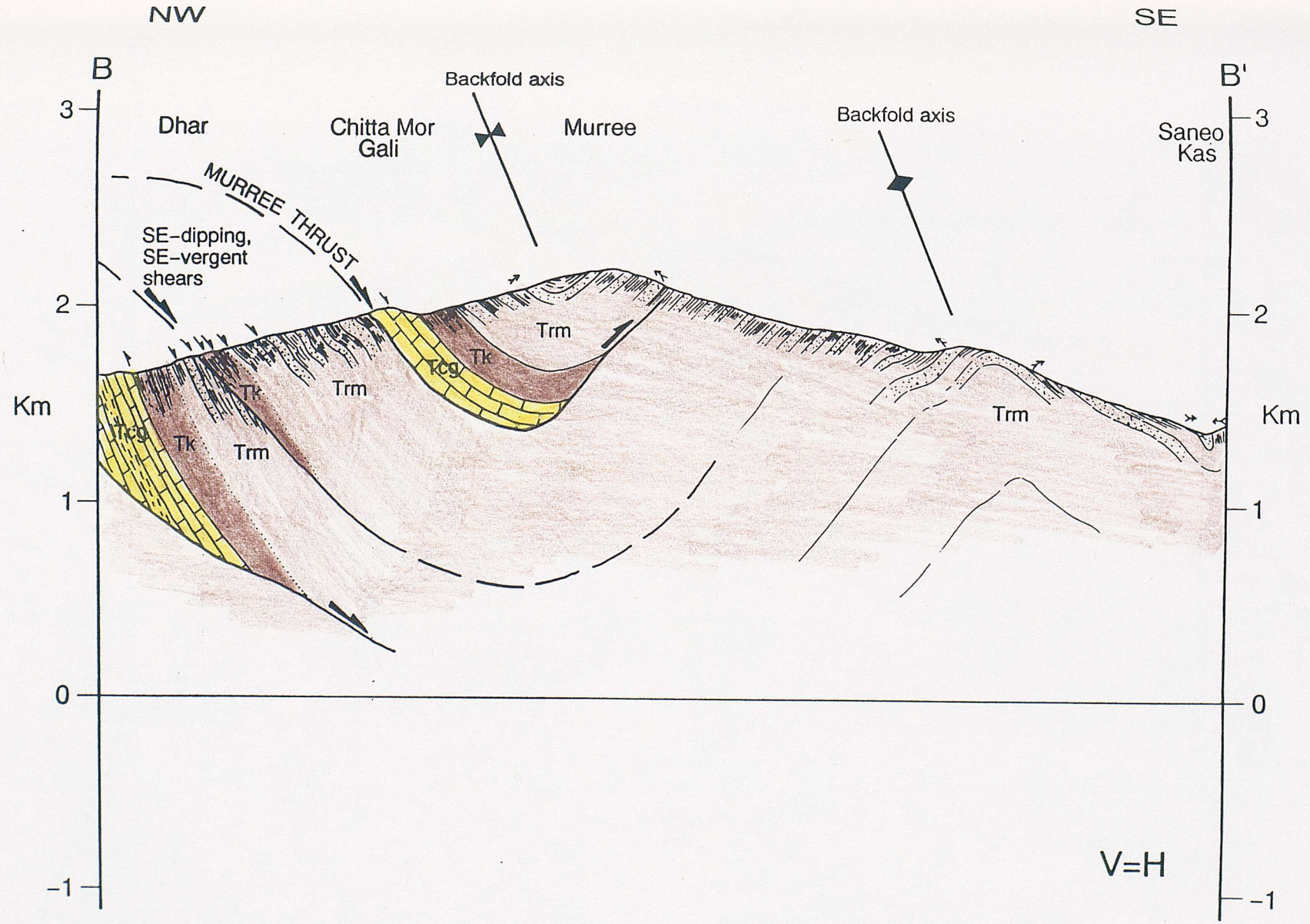


Fig.7.4b. SECTION ACROSS THE MURREE THRUST AT CHITTA MOR GALI/MURREE.

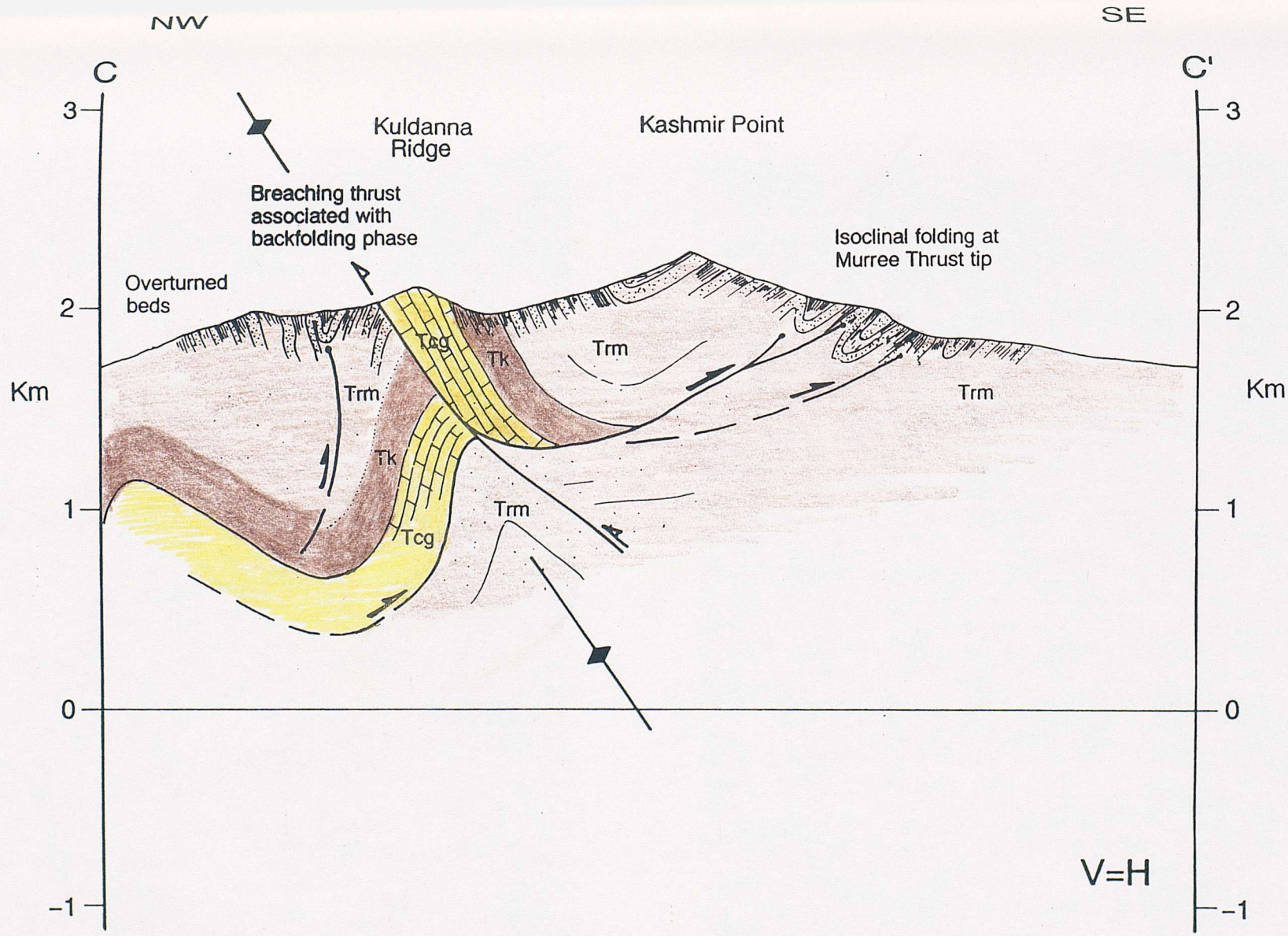


Fig.7.4c. SECTION ACROSS THE MURREE THRUST AT KULDANNA/KASHMIR POINT.





Overturned SE–vergent shears, NW of Bansara Gali

Fig.7.5. SIMPLIFIED STRUCTURAL RE-INTERPRETATION OF LATIF'S (1970) MAP OF S.E. HAZARA.

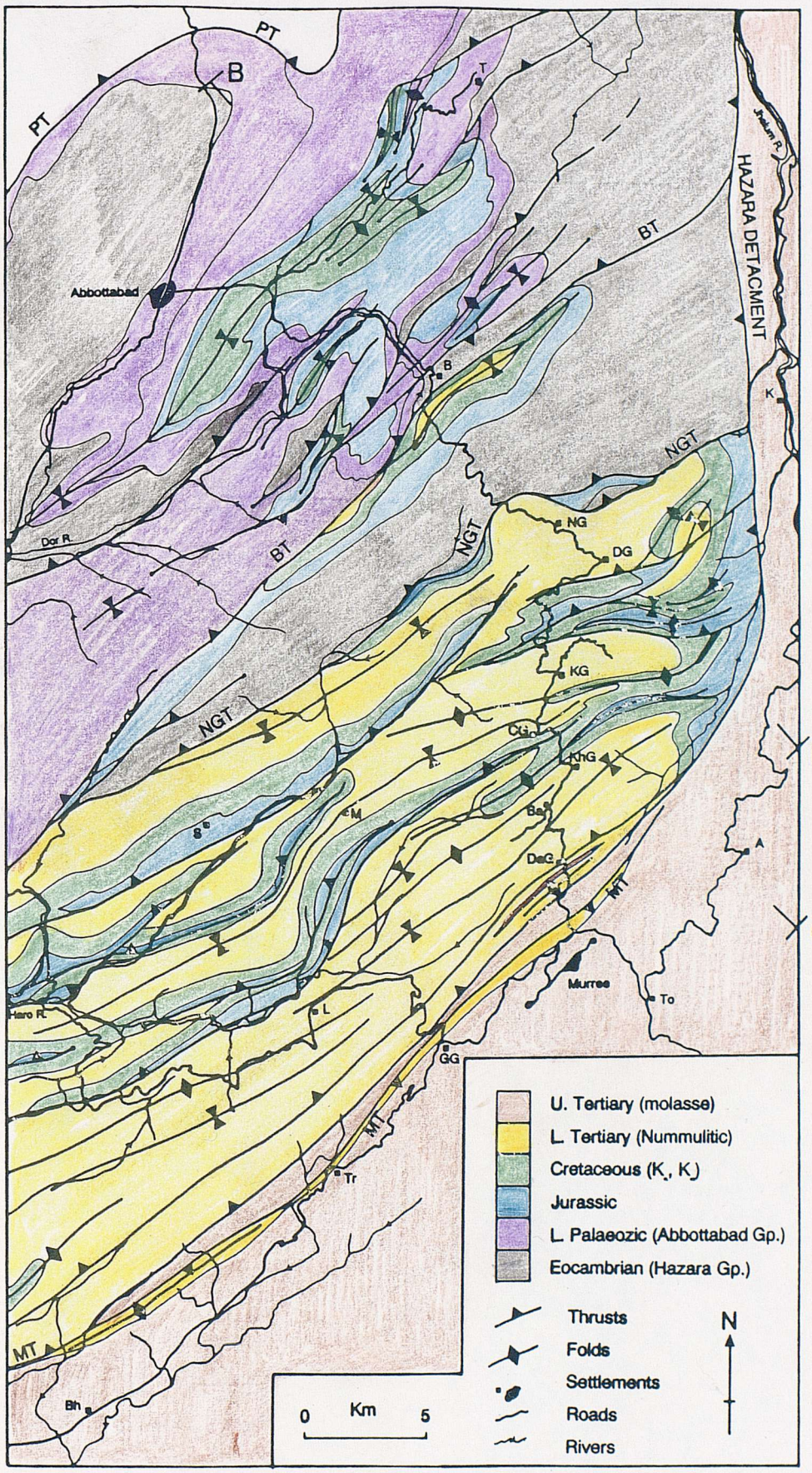


Fig.7.6.

A, Aliot; B, Bagnotar; Ba, Barian; Bh, Bharakao; CG, Changla Gali; DG, Dunga Gali; DaG, Darya Gali; GG, Ghora Gali; K, Kohala; KG, Kuza Gali; KhG, Khaira Gali; L, Lora; M, Mari; NG, Nathia Gali; S, Sataura; T, Thandiani; To, Topa; Tr, Tret

PT, Panjal Thrust; BT, Bagnotar Thrust; NGT, Nathia Gali Thrust; MT, Murree Thrust.

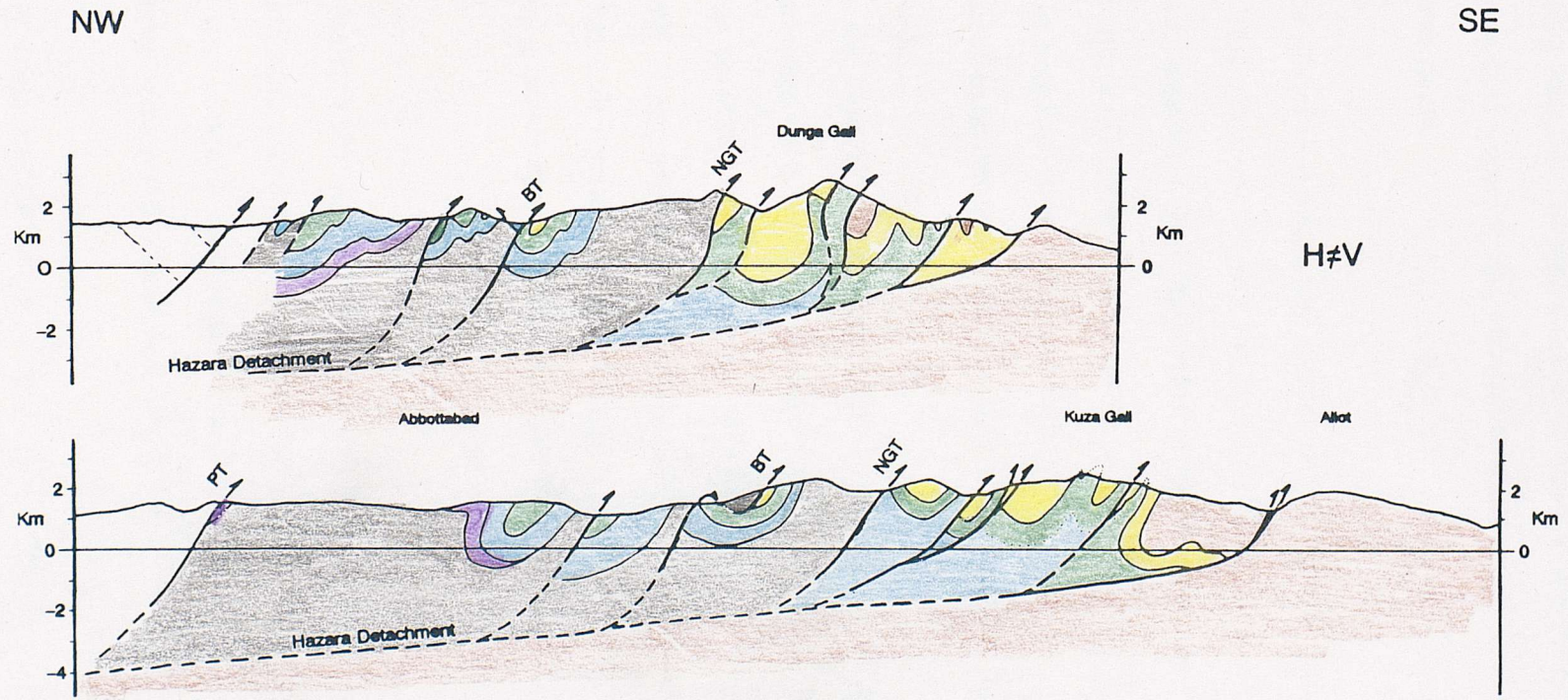


Fig.7.6. SIMPLIFIED STRUCTURAL RE-INTERPRETATION OF LATIF'S (1970) SECTIONS ACROSS S.E. HAZARA.

(ornamentation as accompanying map)

crosses a major synclinal backfold axis and the bedding becomes right-way-up and gently southeasterly dipping. At this point the strata pass from the transitional Kuldanna Formation reddish/purple shales with thin Nummulitic limestones, red sandstones and gypsiferous beds to the Chora Gali Formation which is characterised by platy-nodular, Nummulitic limestones and interbedded greenish marls.

The Chora Gali Formation directly overlies the Murree Formation molasse in a tectonic contact. This contact is the Murree Thrust or MBT. At this point it is emergent, SE-vergent but dips southeastward. Both SE-vergent and NW-vergent shear indicators are observed close to the thrust plane. The NW-vergent shears are interpreted as fold accommodation structures related to the anticlinal axis at the northwestern end of the section, whilst the SE-vergent shears are considered to be indicators of the main transport direction on the Murree Thrust.

In the footwall of the Murree Thrust, below a thin Murree Formation exposure, the Kuldanna Formation outcrops in the core of a second anticlinal backfold. The core has been breached by a late accommodation thrust that may breach the overlying Murree Thrust. At the northwestern limit of the section the bedding is again subvertical to overturned, dipping at 60°SE. Here Chora Gali Formation is again juxtaposed against Murree Formation molasse by a steeply SE-dipping backsteepened thrust. This thrust may be the truncated root of the Murree Thrust.

At depth it is impossible to constrain the structure. Towards the southeast only Murree Formation strata are exposed at the surface. Therefore, thrusts lying in the footwall to the Murree Thrust must comprise only molasse sediments at this point, any Eocene strata having been cut-off against unexposed hangingwall ramps.

A cross-section through Chitta Mor Gali and the centre of Murree itself (Fig.7.4b) shows a much simpler structure, being dominated by open backfold axes around which the early SE-vergent thrust imbricates have been folded. The absence of late breaching

faults associated with the backfolding phase means that the early SE-vergent shearing and imbrication is well-exposed in this section. The Chora Gali, Kuldanna and Murree Formations exposed between Dhar and Chitta Mor Gali comprise right-way-up, SE-dipping imbricates. Their orientation is suggestive of backthrusting, but since shear sense criteria indicate that the imbricates are actually refolded SE-vergent thrusts. In this section the Murree Thrust is therefore a SE-dipping, SE-vergent thrust.

The shear zones exposed between Dhar and Chitta Mor Gali are excellent, but are particularly well-developed in the Kuldanna Formation outcrops where the thinly interbedded evaporites, shales, limestones and sandstones provide excellent markers within a shearing matrix. The shears are dominantly SE-vergent. However, at the extreme northwest of the section NW-vergent shears become more common and this is related to the proximity of the anticlinal back-fold axis which lies beyond the section line. Accommodation shearing is common at the anticli-nal hinges - shearing being towards the fold axis and sometimes culminating in breach thrusting (Figs.7.4a&b).

The kilometric wavelength of the backfolds implies that they are deep-rooted structures. It is proposed that they are related to a NW-vergent roofthrust to a duplex at depth which is discussed later. Detailed mapping has shown that the subvertical attitude of beds throughout the Galiat is not due to passive backsteepening or resistance to motion at the tip of the MBT (Coward & Butler, 1985), but due to the formation of the North Potwar Deformation Zone duplex.

In the easternmost section (Fig.7.4c), the Murree Thrust is not exposed due to a combination of lateral ramping and recent uplift. The lateral hangingwall ramp is observed near Sunny Bank, where the Chora Gali Formation limestones pinch out. The surface limestone exposures in Figure 7.4c., which have in the past (Latif, 1970) been equated with the Murree Thrust, actually lie in the footwall of the Murree Thrust. The Murree Thrust has therefore been eroded or juxtaposes Murree Formation molasse in both its hangingwall and footwall and is

therefore difficult to recognise in the field.

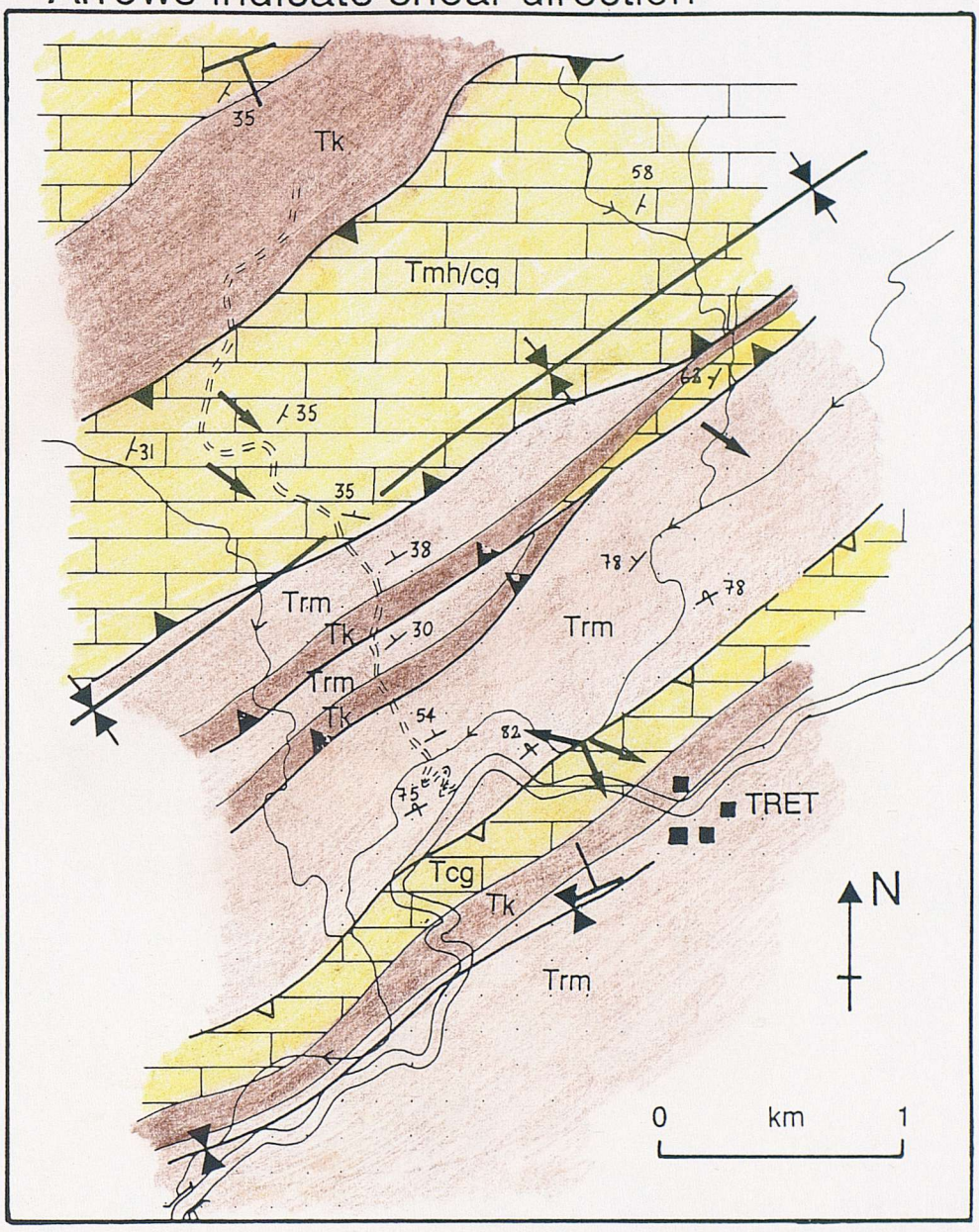
At Kuldanna ridge a thrust plane is exposed. The geometrical relationship between the Chora Gali Formation, in its hangingwall, and the Murree Formation strata that lie in its footwall demands that the thrust is a late breaching structure. The molasse beds at the northwestern end of the section are overturned and young towards the northwest, except for a small-scale isoclinal fold pair. The younging confrontation across the thrust plane dictates that the thrust has breached the hinge of a backfold anticline in a backthrust sense. It is inferred that an early SE-vergent thrust carried Chora Gali Formation over the Murree Formation molasse series but was then folded about a NW-vergent anticlinal axis and, finally, the fold axis was breached by a late accommodation backthrust. Shears within the Chora Gali Formation limestones indicate a SE-vergent shear providing evidence of the early thrust transport direction.

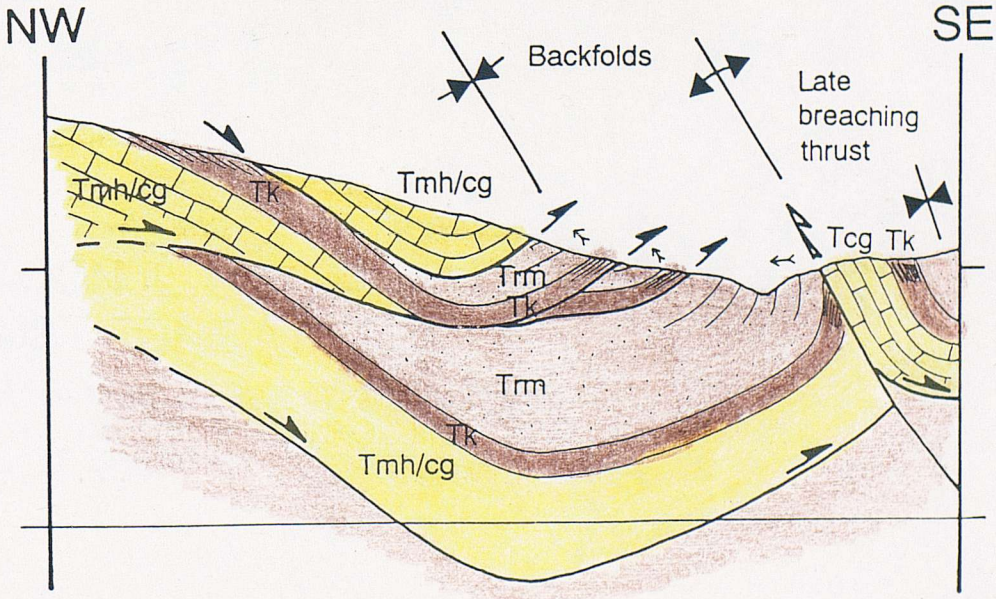
Kashmir Point lies to the southeast of a synclinal backfold axis and thus, the Murree Formation strata observed at outcrop dip moderately (55-65°) northwestward. The younging direction is not consistently northwestward however. Several younging reversals were mapped near Jhikka Gali. The bedding dips lie within a very small range and therefore the folding is isoclinal. Since the folding is localised and faces southeastward these folds are interpreted as tipfolds related to the early phase of SE-vergent thrusting. As mentioned earlier, intramolasse faults are difficult to recognise in the field and so it is possible that some of SE-vergent thrusts are emergent but were not observed.

A further traverse of the MBT zone was mapped at **Tret** which lies on the main Rawalpindi-Murree road approximately 12km SW of Murree (Fig.7.7).

The Murree Thrust at this locality has a breaching backthrust geometry. It is a steeply SE-dipping structure that carries right-way-up Eocene hangingwall strata over inverted Murree Formation molasse in its footwall. The cut-off geometries and breaching style is similar to that of the Murree Thrust at Kuldanna (Fig.7.4c). At the thrust plane, which is well-exposed

Fig.7.7. MAP AND SECTION OF THE MBT ZONE AT TRET, ON THE PINDI-MURREE ROAD.
Arrows indicate shear direction





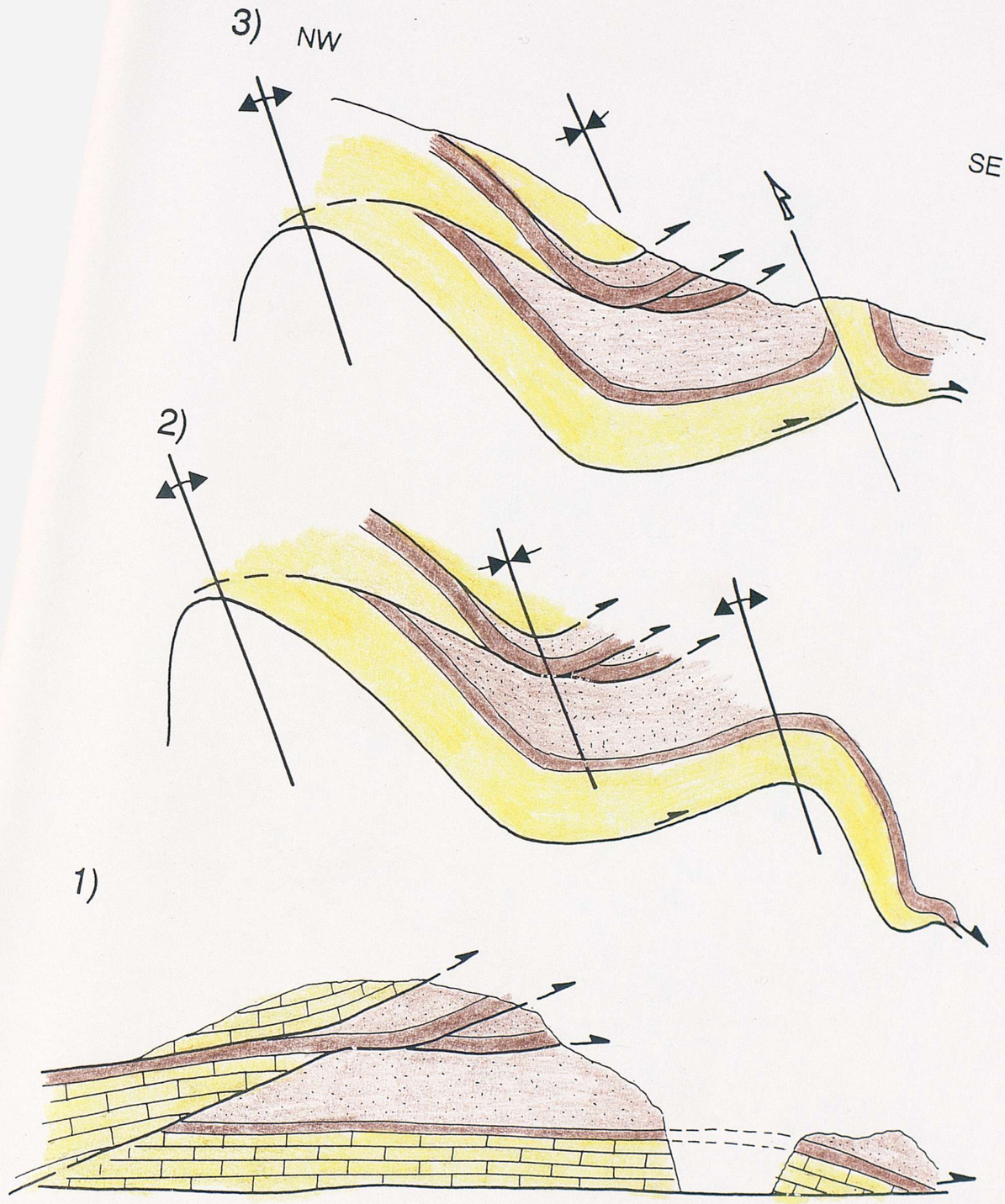


Fig.7.8. CARTOON ILLUSTRATING THE EVOLUTION OF THE MBT ZONE AT TRET.

1) EARLY SE-VERGENT IMBRICATION
2) NW-VERGENT BACKFOLDING
3) LATE BREACHING OF ANTICLINAL AXIS

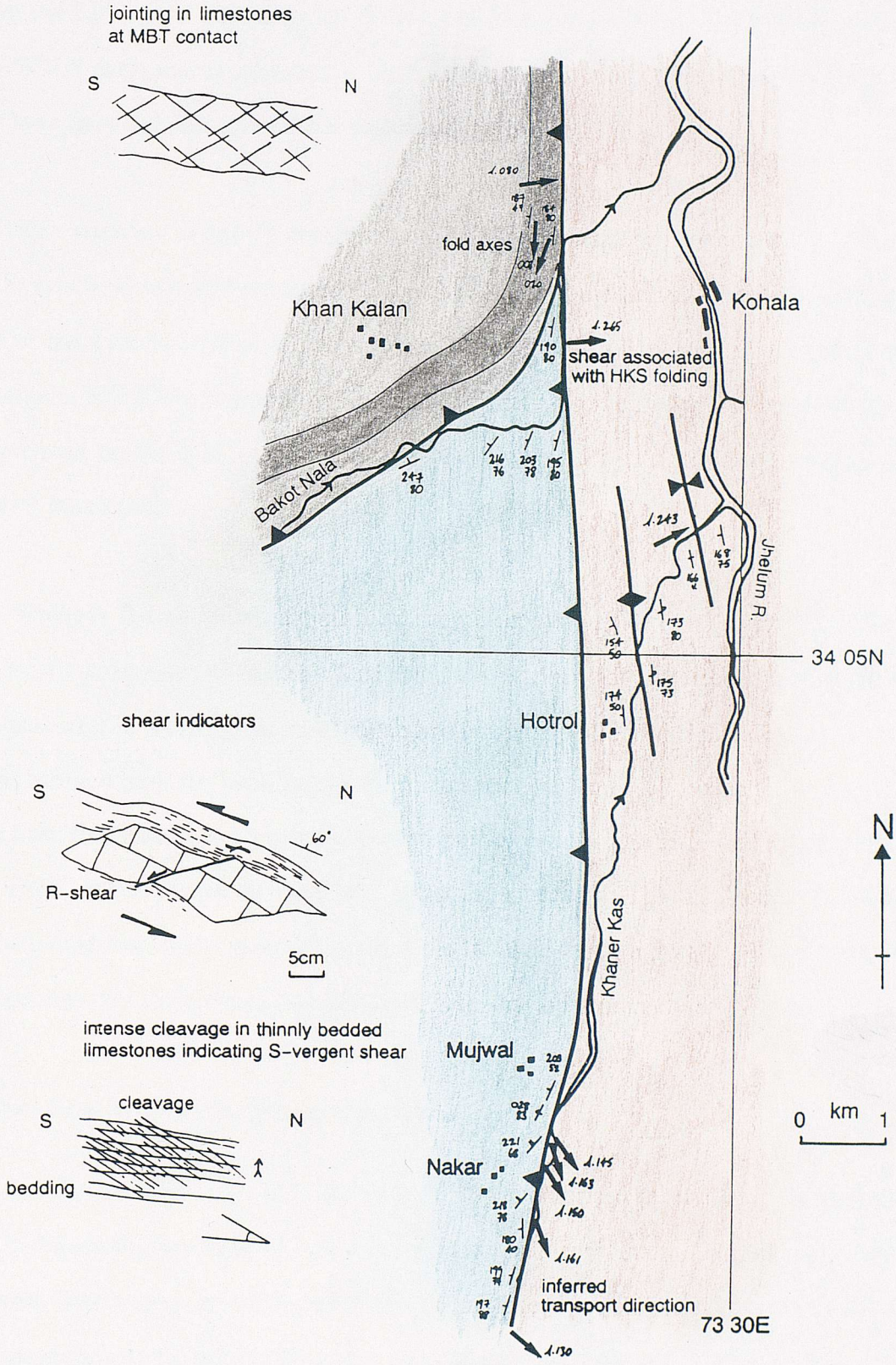


Fig.7.8b STRUCTURAL DATA FROM THE HAZARA DETACHMENT (MBT) AT KOHALA.

in the river section NW of Tret, thin gypsiferous bands are found in the highly sheared shales suggesting that the Kuldanna Formation lies in the immediate subsurface. Lineations indicate an E–W to ESE–WNW shear that is restricted to the breaching thrust alone. The imbricated strata to the northwest have been sheared towards the southeast (N126E–N136E).

The structure (Fig.7.7) comprises an imbricated sequence that has been folded around a major synclinal axis and subsequently breached during the last stages of backfolding. The evolution of the zone is illustrated by a series of cartoons (Fig.7.8). During imbrication the base of the Margalla Hill Formation and the Kuldanna Formation served as detachment horizons. Further displacement on the MBT zone imbricates was prevented by backfolding, and finally, breaching thrusts developed.

Towards the northeast, along the western bank of the Jhelum River, the Hazara Detachment (MBT) is exposed (Fig.7.8b). The intense shearing that occurred on the Hazara Detachment is manifest in the different rocks in the MBT hangingwall. The Hazara Group develops a mylonitic fabric whilst the Samana Suk Formaton becomes intensely veined and cleaved. Lineation data from the detachment constrain the transport direction to N150°–165°E, although a second set of E–W shears related to HKS folding is also observed. The displacement on the MBT can be estimated from the minimum length of the footwall–hangingwall cut–off normal to the transport direction. On the western limb of the HKS this is approximately 85–90km.

7.1.2. Kuldanna–Khaira Gali : the Nummulitic Zone.

The Nummulitic Zone of Middlemiss (1896) comprised all the country between Kuldanna and Nathia Gali. Here I refer only to that region between Khaira Gali and Kuldanna where the Hazara Detachment or MBT, which forms the sole thrust to the Lesser Himalayan Zone, has cut up–section so as to lie within the Lower Tertiary succession. The region therefore represents the deformation tip of the MBT wedge. The MBT did not emerge in the Murree area, or anywhere along the Margalla Hills or Kallachitta Hills mountain front. Rather, the MBT propagated within the Murree Formation molasse to the south.

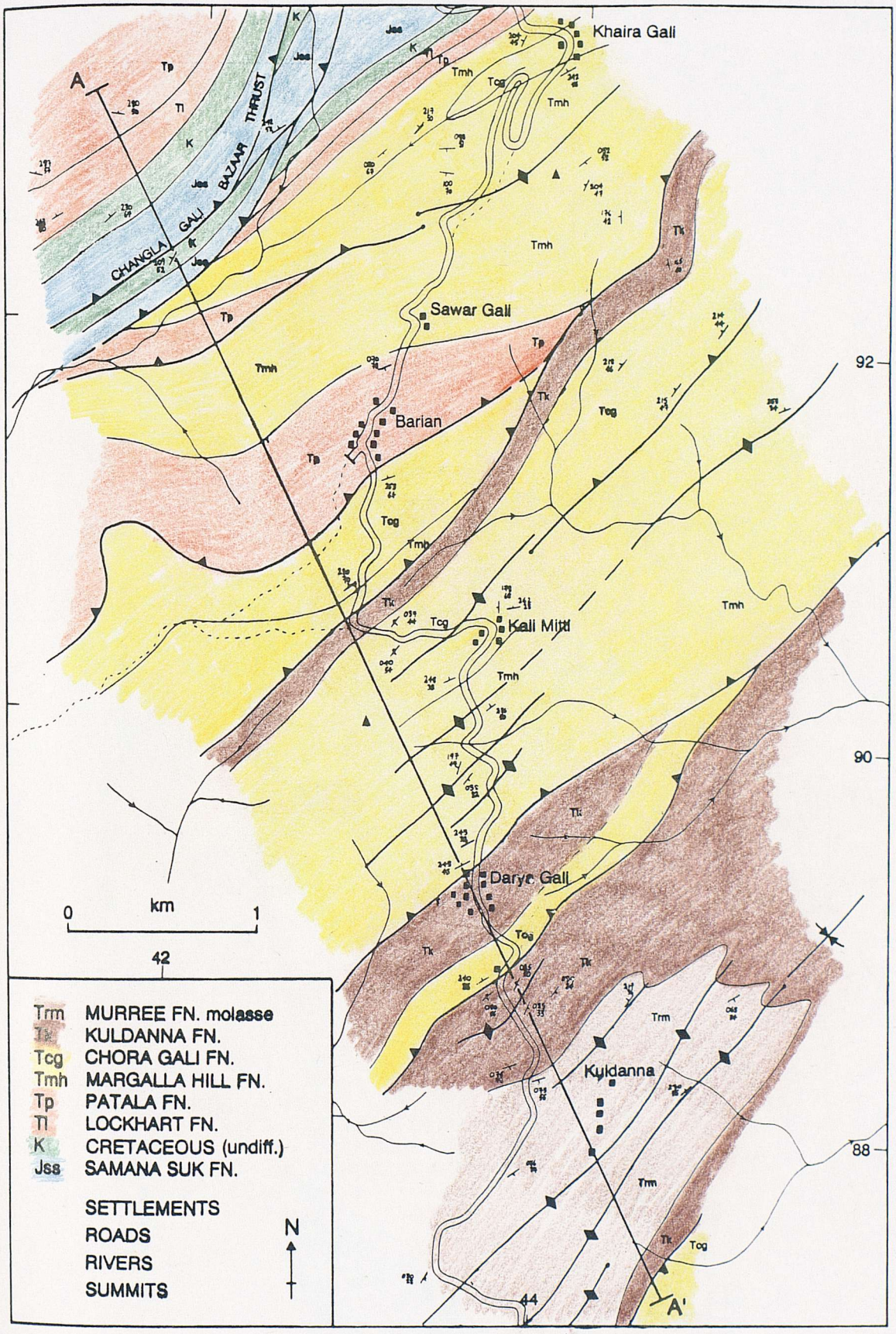


Fig.7.9. MAP OF THE NUMMULITIC ZONE BETWEEN KHAIRA GALI AND KULDANNA.
Section A–A' is shown in Figure 7.10

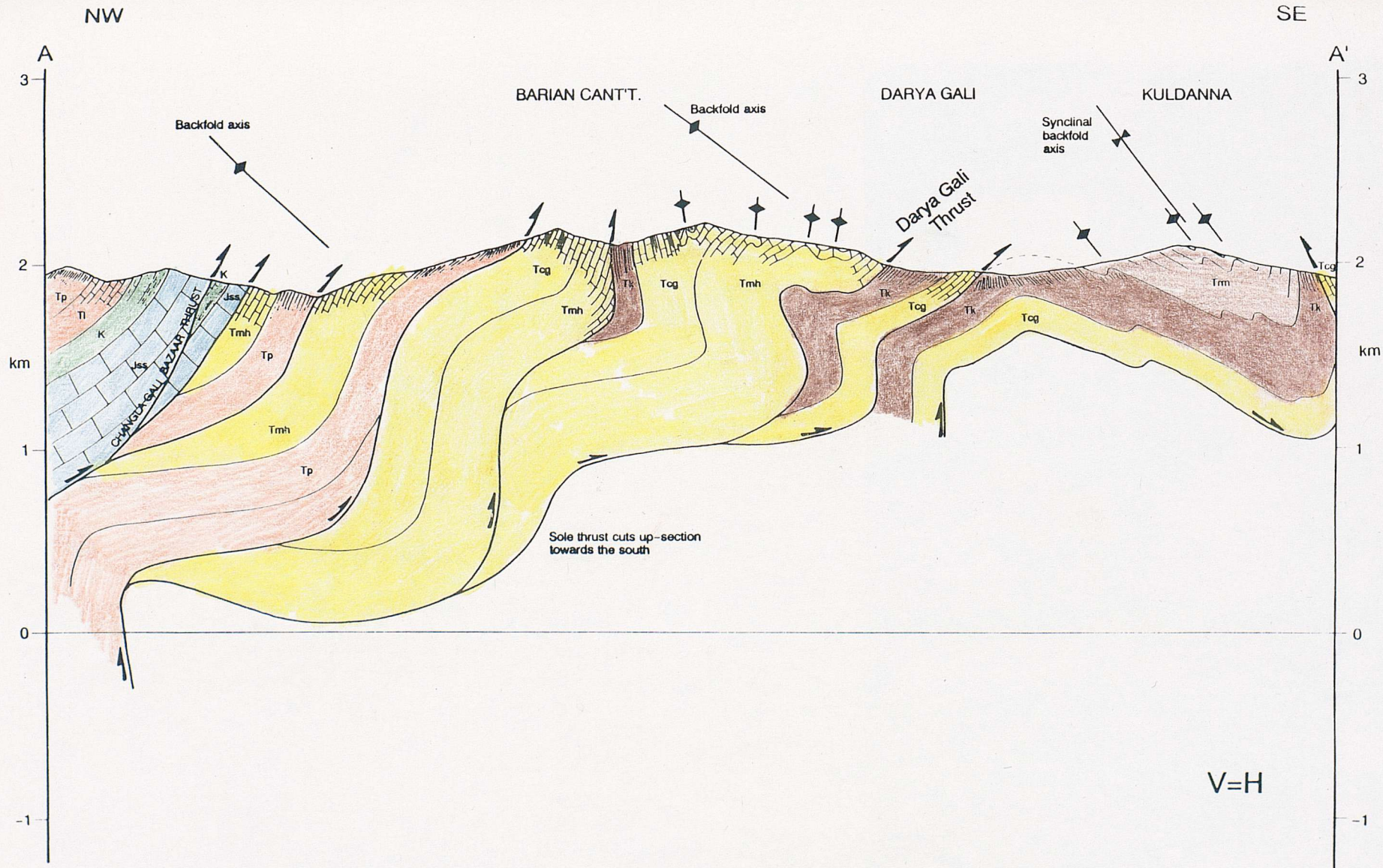
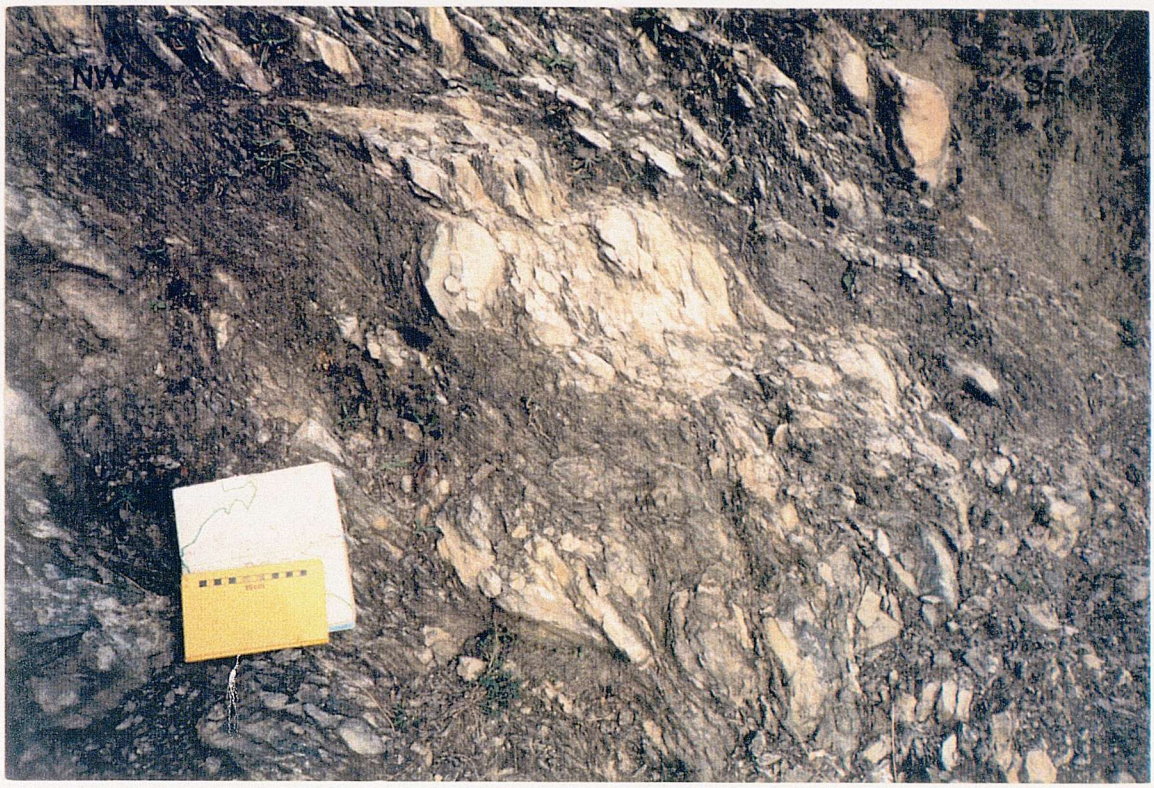
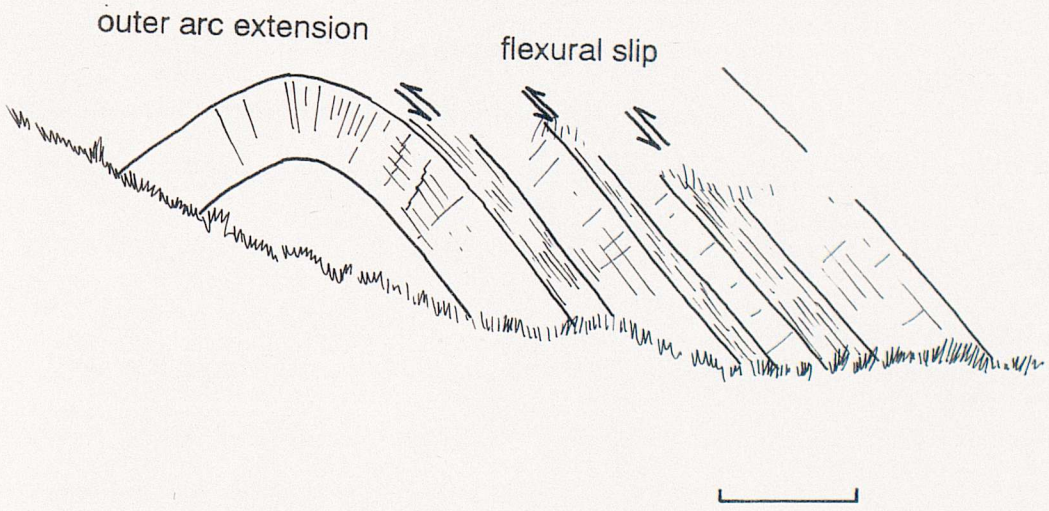


Fig.7.10. SECTION THROUGH THE NUMMULITIC ZONE BETWEEN KHAIRA GALI AND KULDANNA.



(a) Shearing of the Chora Gali Formation



(b) Folding of the interbedded limestones and shales of the Margalla Hill Formation

Kuldanna lies only 2km from Murree at the beginning of the Murree–Abbottabad road (Fig.7.9). Between Kuldanna and Khaira Gali only Tertiary strata are exposed, belonging to either the Palaeocene–Eocene Galis Group or the Miocene Rawalpindi Group. The former comprises Nummulitic limestones and shales that were deposited on the Indian shelf just prior to closure of Neotethys. Marine conditions persisted for some time after India–Asia collision. The Kuldanna Formation represents a transitional unit between marine and continental conditions, marking the onset of terrigenous input. It has been assumed that the Oligocene was a period of non–deposition prior to the onset of fluvial molasse deposition in Miocene times, represented by the Rawalpindi Group. Bossart & Ottiger (1990) report marine sedimentary environments at the base of the Rawalpindi Group.

Because no Mesozoic strata are exposed it is inferred that the Hazara Detachment or MBT had ramped up to Palaeocene and eventually Eocene horizons in this area. Fieldwork in Khaner Kas, where the Hazara Detachment is exposed to the northeast, of Kuldanna confirmed.

The Kuldanna area lies in the core of a large scale synclinal backfold (Fig.7.10). Murree Formation molasse is preserved in the fold core and as one moves away from the axis the detrital Fatehjang Member is observed on both fold limbs. Small scale parasitic folds have been mapped in the fold core. At the southeastern limit of Figure 7.10, the overturned fold limb has been breached by a late thrust associated with fold hinge accommodation (Fig.7.4c). This thrust juxtaposes Chora Gali Formation against either the Murree (Fig.7.4c) or Kuldanna Formations (Fig.7.10).

On the right–way–up limb of this syncline extensive exposures of Kuldanna Formation occur. These outcrops are intensely sheared, shear sense indicators showing both southeast– and northwest–vergent shearing. In places both of these shear senses are observed at the same locality suggesting that the true deformation is in fact a bedding–parallel compression. This is in accordance with the large scale structural geometry of this region, since an anticlinal

backfold axis lies to the south of Darya Gali. The foliation of the Kuldanna Formation shales fans between 40°SE and 70°NW, passing through subvertical orientations. This is interpreted as a late backfold-related foliation.

Evidence of early SSE-vergent thrusting is exposed just south of Darya Gali. Here two imbricates carry Chora Gali and Margalla Hill Formation limestones over Kuldanna Formation shales. The structurally lower imbricate is relatively thin, but a full Eocene sequence is found in the second imbricate - the Darya Gali Thrust. The Eocene succession approaches 1000m in thickness in this area, although tectonic thickening may have occurred - which contrasts markedly with the sequence preserved in the Salt Ranges to the south which is only 100-200m thick. The folds in the hangingwall of the Darya Gali Thrust are upright to SE-vergent. The Darya Gali Thrust sheet has deformed by flexural slip within the interbedded limestones and shales of the Eocene succession. Kuldanna Formation red shales and sandstones are preserved in the hangingwall of the Darya Gali Thrust and are not exposed again until near Nathia Gali.

The section between Barian and the Changla Gali Bazaar Thrust is typified by bedding dips that fan between moderate northwesterly dips and subvertical to overturned southeasterly dips. This is the result of flexure around gently SE-dipping, NE-SW striking fold axial planes. The attitude of bedding is also contributed to by backsteepening during the stacking of thrust imbricates.

At Khaira Gali an inverted stratigraphic sequence from Chora Gali to Samana Suk Formation is observed. These strata are tightly folded around a SE-vergent tipfold that at depth is represented by a thrust. The root of this thrust is shown in the footwall of the Changla Gali Bazaar Thrust (Fig.7.12a).

Evidence of late SW-NE directed compression, which is well-developed in Ayubia is sparse in the Nummulitic Zone. However, swings in strike are mapped on the ridge descending southeastwards from Khaira Gali indicating some late crumpling of this region. This

deformation phase is described more fully in section 7.1.4.

7.1.3. Khaira Gali–Kuza Gali : the environs of Changla Gali

Changla Gali, which is situated 16km from Murree on the Murree–Abbottabad road, stands at almost 2700m. The road descends northwards to Kuza Gali and southwards to Khaira Gali, which lie 19km and 12 km from Murree respectively (Fig.7.11). The road lies on the Jhelum–Indus watershed and altitude is lost rapidly to both east and west.

To the north of Khaira Gali one encounters Mesozoic strata for the first time in the Galiat – testifying to the northerly downcutting of the Himalayan thrust wedge. The cross–sections through the area are dominated by Jurassic and Cretaceous strata; the Tertiary succes–sion being confined to the uppermost 500–1000m.

It is in the environs of Changla Gali that one observes a major change in the upper Cretaceous stratigraphy. To the south of Changla Gali the entire Cretaceous succession totals only 100–150m. This thickness comprises primarily shales and ferruginous sands of the Chichali and Lumshiwal Formations; the upper Cretaceous being represented by a 20–30m thick soft–weathering, marly limestone. The Hangu Formation, representing the uppermost Cretaceous to basal Palaeocene period, comprises a well–developed pisolitic laterite indicating emergence and subaerial erosion prior to deposition of the lower Tertiary Nummulitic limestone series.

To the north of Changla Gali the upper Cretaceous succesion is represented by a 500m thick well–bedded, hard–weathering, white micritic limestone. The basal portion of the Kawagarh Formation in this area is represented by a softer, greenish, marly limestone which may be equivalent to the entire upper Cretaceous sequence preserved to the south. The Hangu Forma–tion is a more poorly–developed laterite than observed further south.

A sedimentological problem therefore remains as to whether the ground to the south of Changla Gali was emergent for much of the upper Cretaceous or whether end Creta–

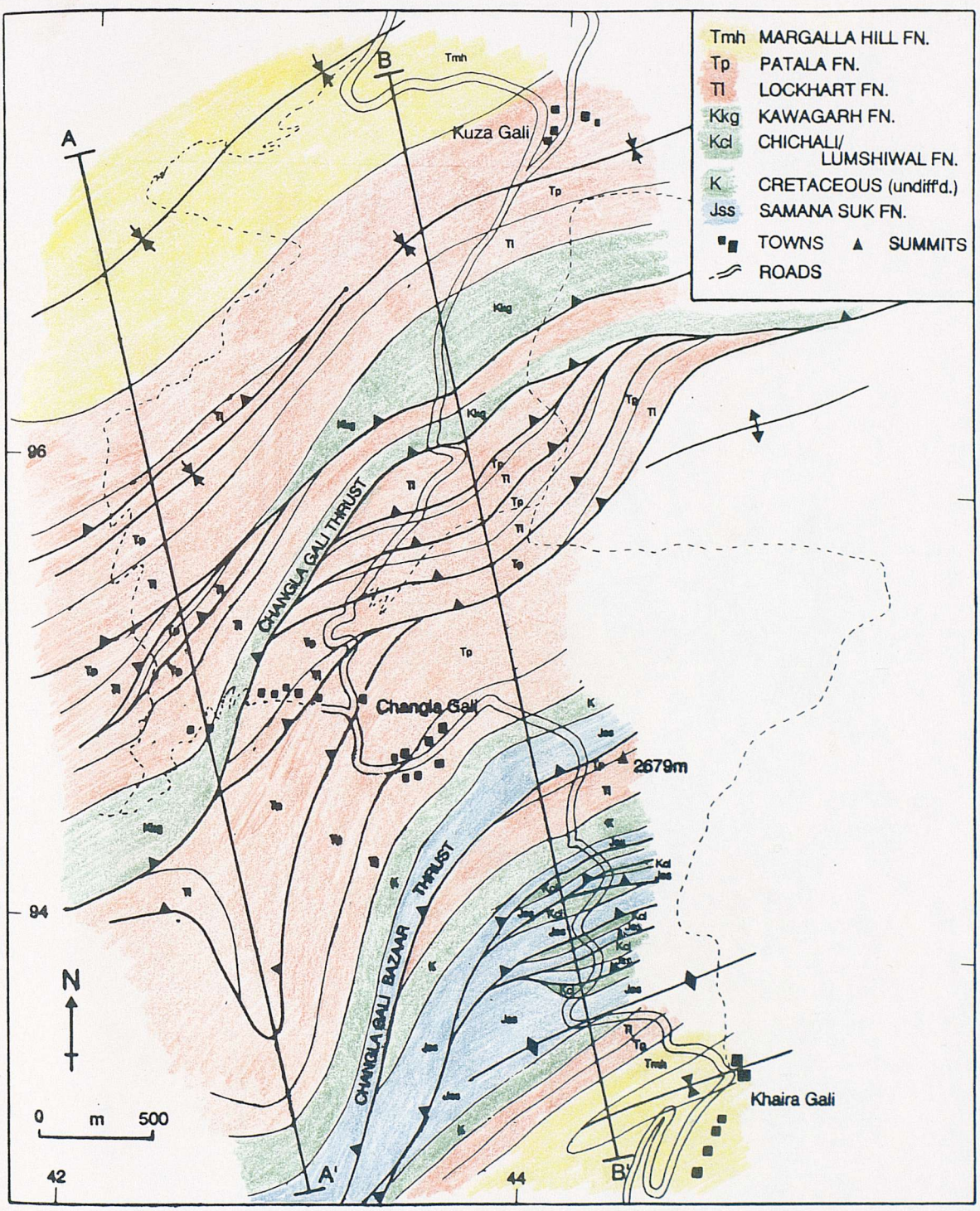


Fig.7.11. MAP OF THE AREA BETWEEN KUZA GALI AND KHAIRA GALI.

Sections A–A' and B–B' are shown in Figures 7.12a&b respectively.

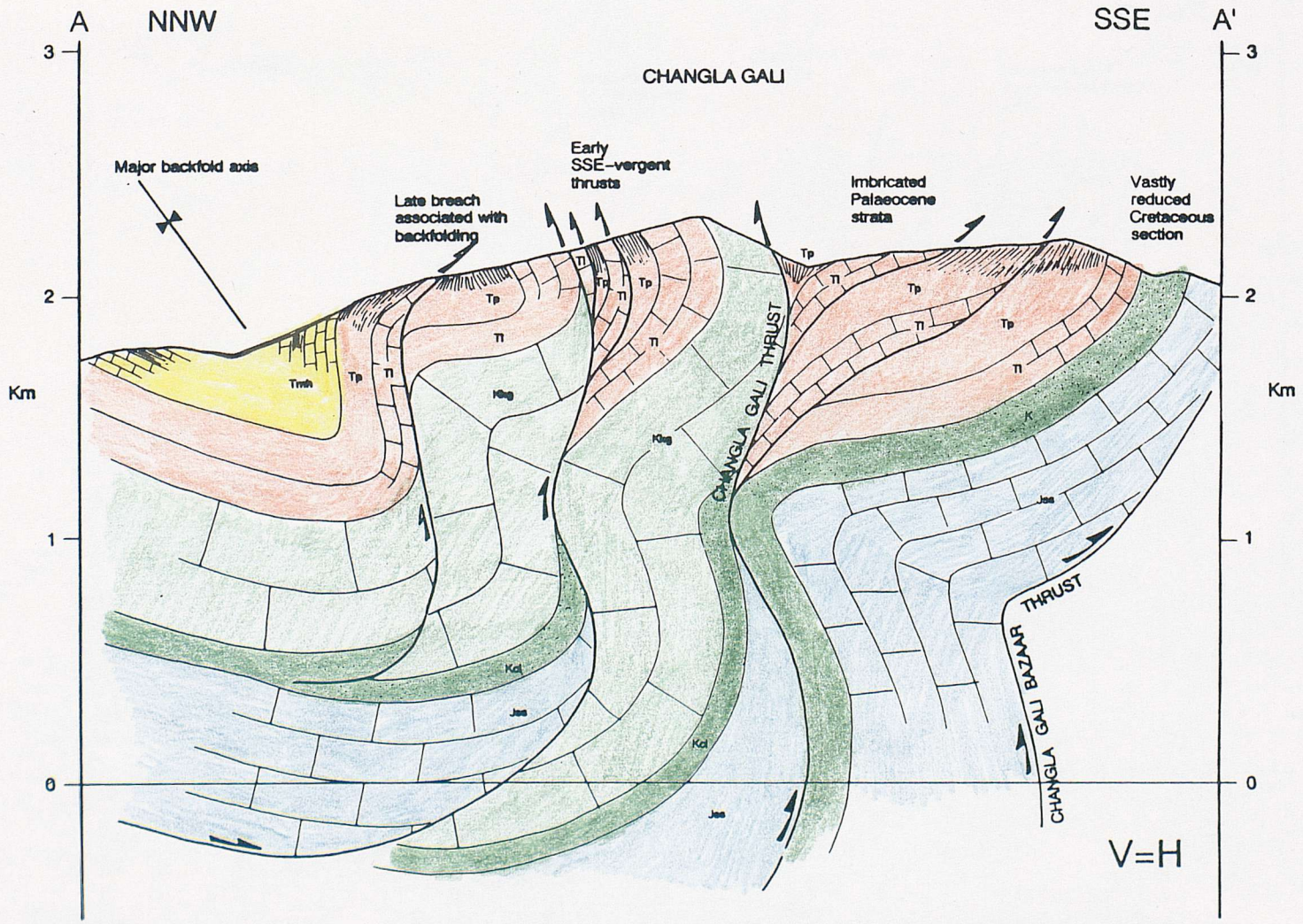


Fig.7.12a. SECTION THROUGH CHANGLA GALI, WEST OF THE MAIN ROAD.

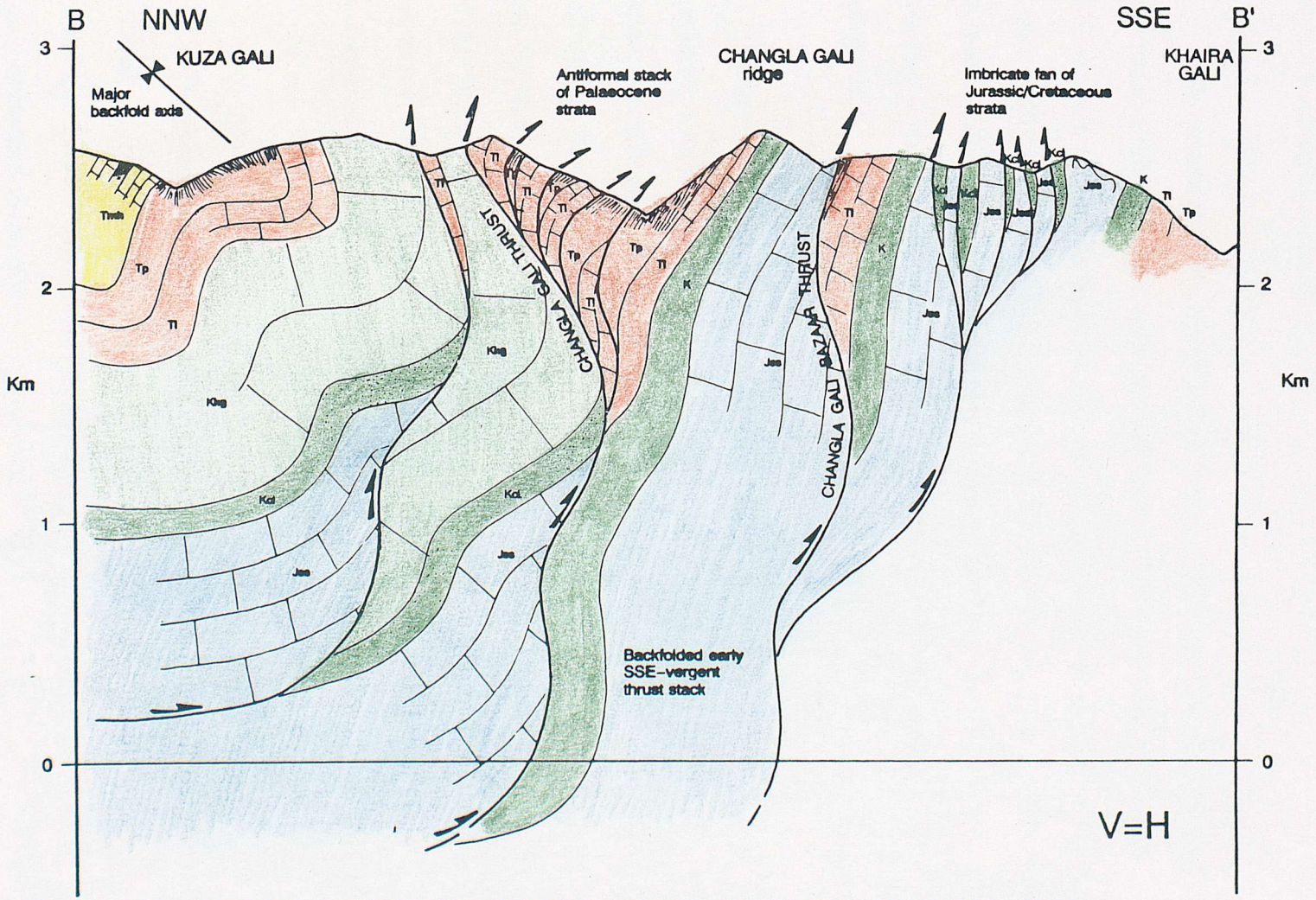


Fig.7.12b. ROAD SECTION FROM KUZA GALI TO KHAIRA GALI.

ceous uplift was greater towards the south. Equally, upper Cretaceous extension may be responsible for facies and thickness variations, but no normal faults have been observed in the field. This could be a function of exposure.

The structure of the Changla Gali region is complex due to the preservation of Palaeocene strata in this area. The Lockhart and Patala Formations tend to decouple from the Kawagarh Formation and imbricate in the footwall of major over-riding thrusts to form imbricate fans and antiformal stacks. The greater complexity of the original SSE-vergent thrust geometry tends to disguise the backfold axes that are so clearly observed, to the north, in Ayubia. The backfolding is manifest upon descent into the valleys when the fanning of bedding dips becomes diagnostic.

Figure 7.12a is characterised by three significant SSE-vergent thrust slices involving the Jurassic to lower Tertiary succession. The most southerly of these, the Changla Gali Bazaar Thrust, becomes emergent just to the south of the section where it carries the Samana Suk Formation over the Kawagarh Formation. The thrust ramps laterally towards the east such that the Jurassic overlies the Palaeocene Patala Formation (Fig.7.11). As it ramps up-section, a number of thin imbricates comprising the uppermost Samana Suk Formation and the attenuated Cretacoeus series formed in the footwall to the Changla Gali Bazaar Thrust. The thin, shale-dominated nature of the Cretaceous sequence in this area means that it is exploited as an easy slip horizon.

The Palaeocene strata in the hangingwall of the Changla Gali Bazaar Thrust have been imbricated to form an antiformal stack. In Figure 7.12b this is particularly well developed, but in Figure 7.12a only the basal portion of this stack is preserved and the imbricates are moderately-dipping compared to the sub-vertical beds of Figure 7.12b. A minimum of 4-5km displacement has occurred on the Changla Gali Thrust. At depth this is accommodated on a single thrust plane, with Jurassic hangingwall strata being carried along an upper Cretaceous footwall flat.

Importantly, **the Changla Gali Thrust is the first thrust to incorporate a thick upper Cretaceous sequence in its hangingwall**. It is possible therefore that displacement on this thrust may be greater than that shown (Figs.7.12a&b).

The thrust in the hangingwall to the Changla Gali Thrust imbricate (comprising the Jurassic to lower Tertiary sequence) accommodates 1.5–2km shortening (Figs.7.12a&b). To the east of section B–B' (Fig.7.12b) this thrust branches with the Changla Gali Thrust at an intra–Jurassic horizon. Towards the west one can trace the thrust plane as it ramps laterally up–section until it incorporates Palaeocene rocks in its footwall, at which point Palaeocene imbricates develop.

Eocene strata in the hangingwall of the Changla Gali thrust are exposed just north of Kuza Gali. The thrust is strongly folded about a NNW–vergent syncline–anticline pair at this point. In response to this backfolding an out–of–syncline thrust of small displacement (200–300m) was formed.

7.1.4. Kuza Gali–Dunga Gali : Ayubia National Park

The Ayubia National Park, named after one of Pakistan's past presidents, occupies the high ridge between Kuza Gali and Dunga Gali on the Murree–Abbottabad road. Kuza Gali lies 19km and Dunga Gali lies 29km from Murree (Fig.7.13). Although the main entrance to the park is via Khanspur Cantonment, access is also possible from Dunga Gali using the old road which is now a footpath. The attraction of a chair lift makes Ayubia a popular tourist destination.

The peaks within Ayubia reach 2700m, whilst Moshpuri to the northwest is 2822m high. Combined with the fact that the Haro River has incised down to about 1000m above sea–level, this means that there is almost 2km of relief in this area. Since the roads that traverse this area are relatively flat and follow the ridges it is tempting for the geologist to rely on road sections alone. However, it is only upon descent into the valleys that the true geological structure

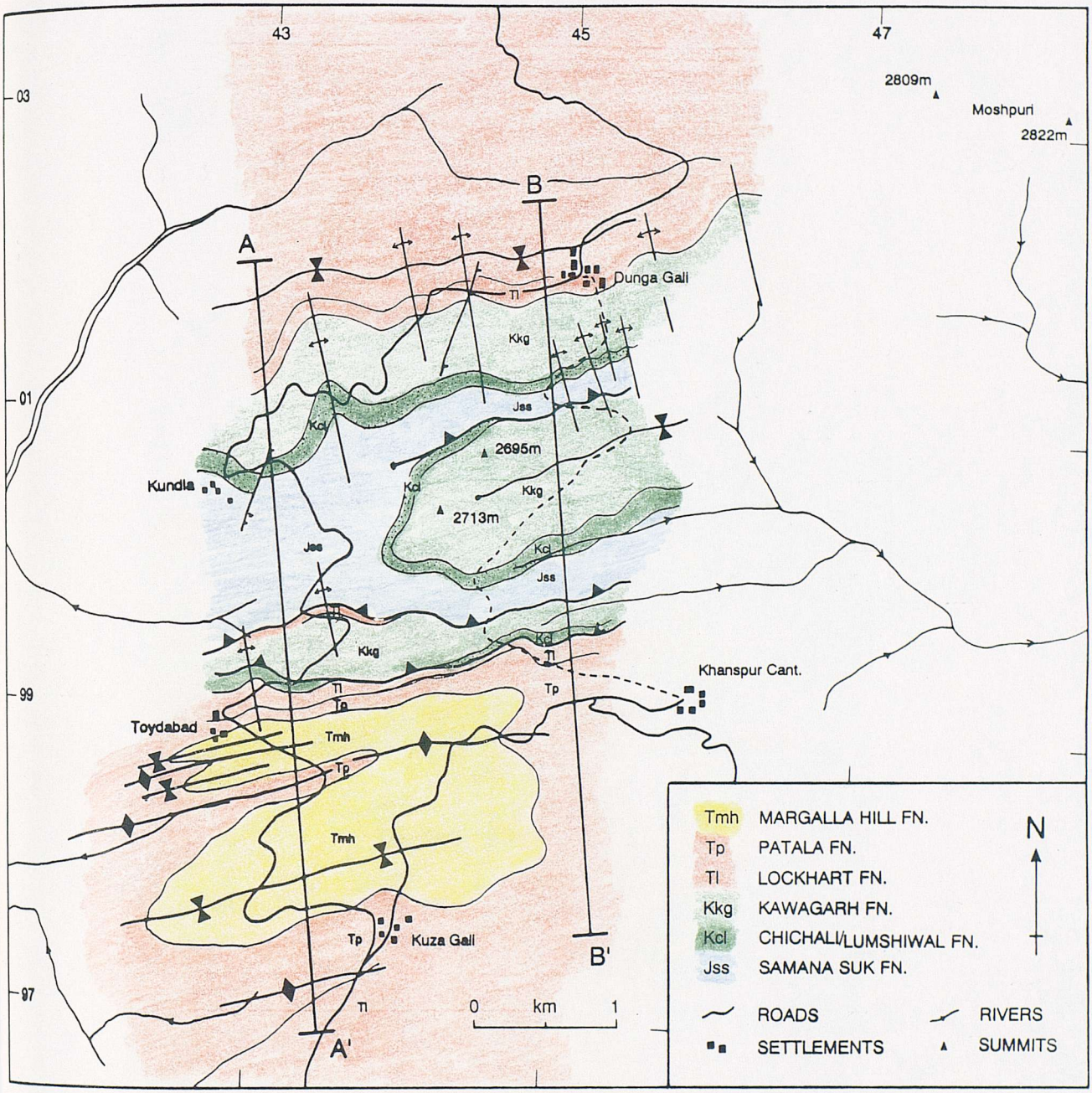


Fig.7.13. MAP OF DUNGA GALI AND AYUBIA.
Sections A–A' and B–B' are shown in Figures 7.14a&b respectively.

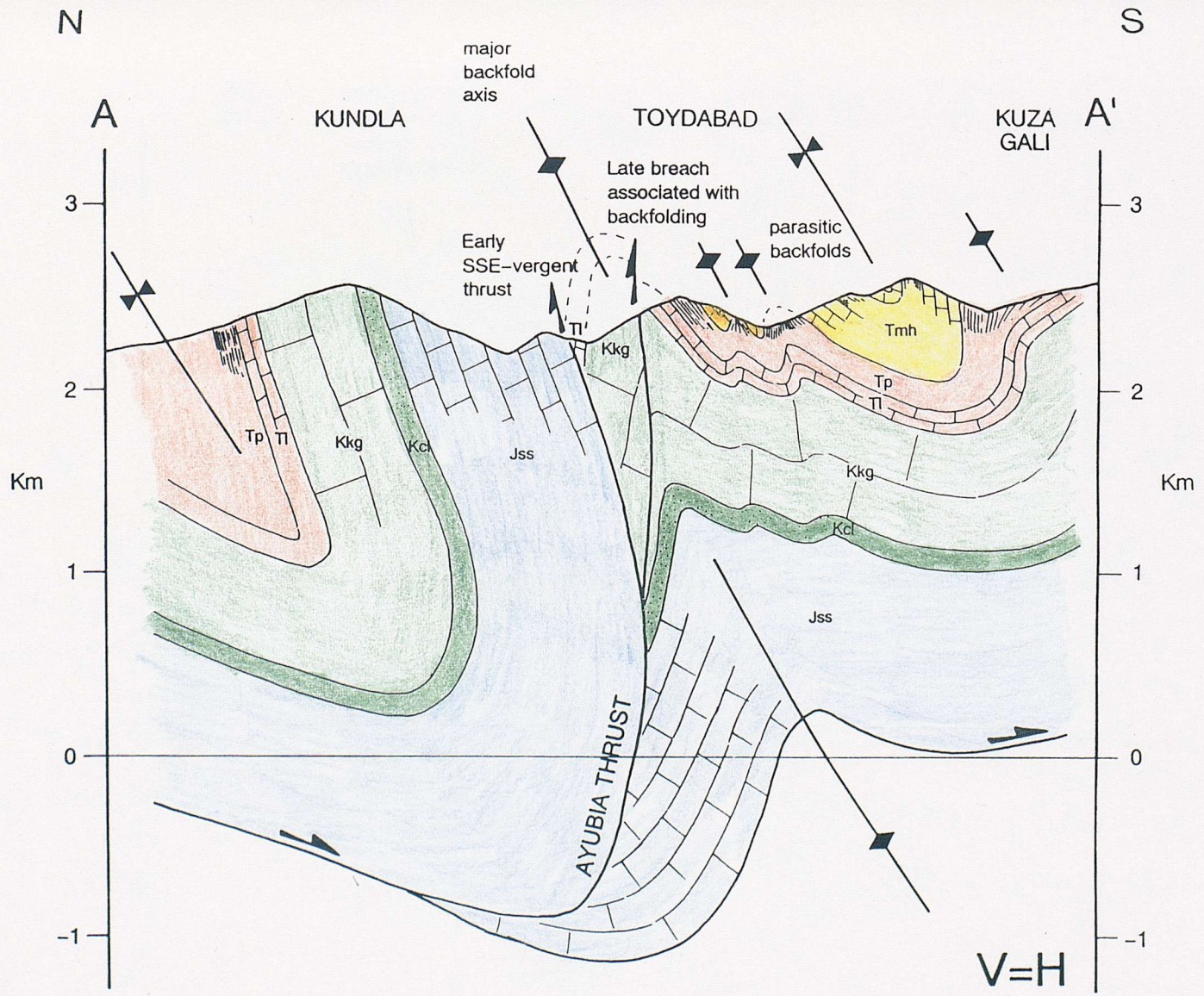


Fig.7.14a. ROAD SECTION FROM KUNDLA TO KUZA GALI.

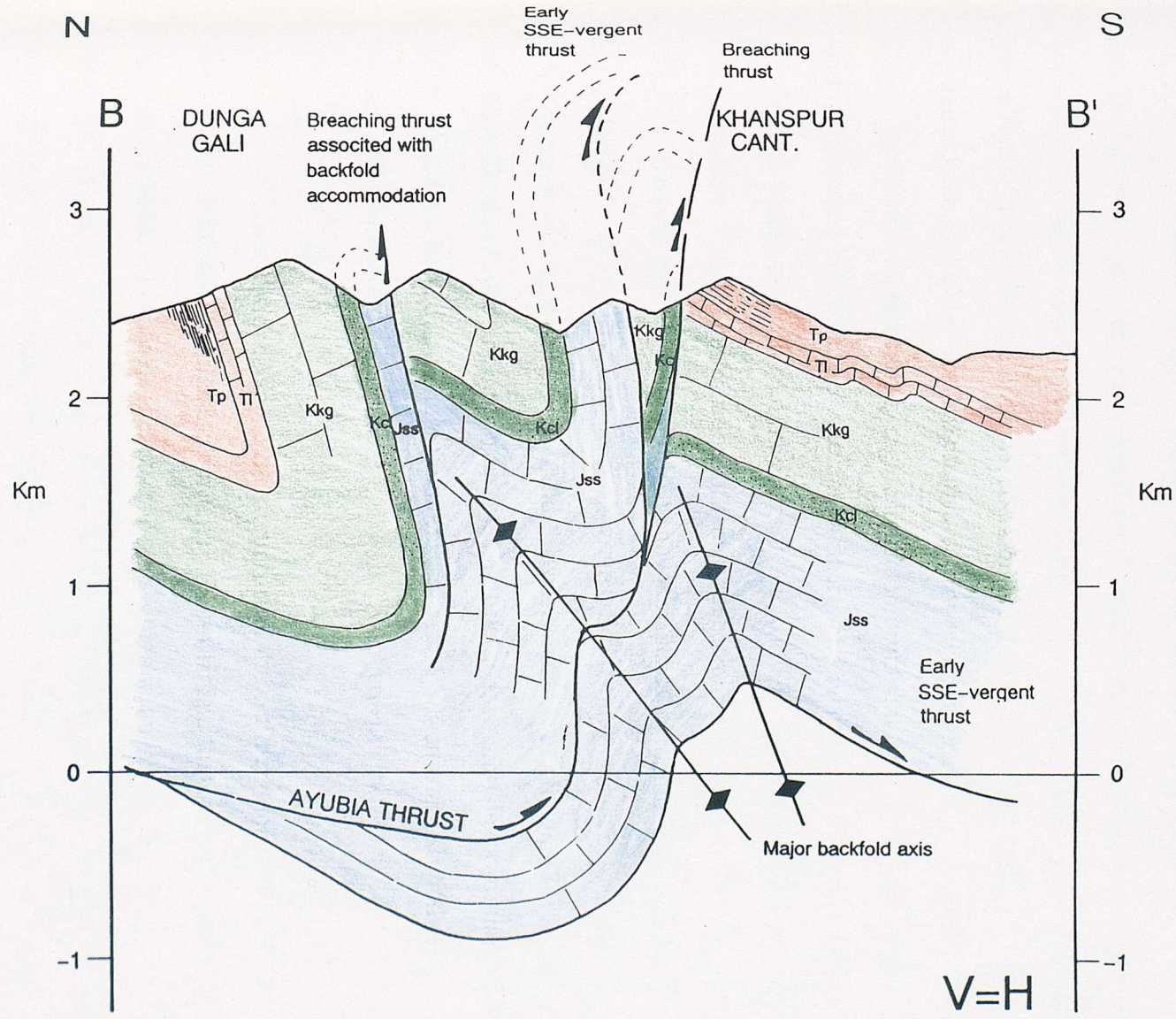


Fig.7.14b.SECTION THROUGH AYUBIA NATIONAL PARK FROM DUNGA GALI TO KHANSPUR CANTONMENT.

becomes obvious.

Ayubia lies in the footwall of the Nathia Gali Thrust (NGT) and so only Mesozoic and Tertiary strata are exposed. All members of the stratigraphic sequence between the Samana Suk to the Margalla Hill Formations are found. Importantly, the upper Cretaceous Kawagarh Formation is very well developed, reaching 550m in thickness. This thickness is estimated from exposures near Kundla (Figs.7.13 & 7.14a) but tectonic thickening may have occurred. The Hangu Formation, representing regional unconformity at end Cretaceous times is therefore poorly developed compared to outcrops further south where major portions of the stratigraphy are omitted. Distinction between the Samana Suk and Kawagarh Formations in the field is difficult, especially since the imbrication of the sequence renders reference to adjacent units unreliable. The Samana Suk Formation often shows evidence of a high energy environment such as oolitic horizons as well as detrital sand grains. Sedimentary structures are also common. The Kawagarh Formation, on the other hand, is consistently a very fine-grained, recrystallised micrite devoid of sedimentary structures. Its basal portion, which is exposed immediately east of Kundla, is characterised by a highly bioturbated surface which is a rusty red colour due to iron-staining and contains pyrite nodules.

As mentioned above **the structure of the Galiat is dominated by the superposition of two deformation phases; an early SSE-vergent thrust imbrication and a NNW-vergent backfolding phase**. The style of this deformation is dictated by the *c.*1500m thick Mesozoic sequence, comprising major carbonate units (Samana Suk/Kawagarh Fns.) separated by a thin shale and sandstone unit (Chichali/Lumshiwal Fns.). This thick competent unit formed large thrust imbricates that comprised the Jurassic-Lower Tertiary sequence. The Lower Tertiary units were imbricated and folded at a small scale in the footwall of the major Mesozoic imbricates due to the presence of abundant shaley units.

In Ayubia, however, a late cross-folding phase is also recognised. It developed in response to NE-SW compression related to the Kashmiri thrust wedge.

The Changla Gali Thrust imbricate and the Ayubia Thrust imbricate are both exposed in this region (Figs.7.13, 7.14a&b). The Ayubia Thrust loses displacement from 4km in the west (Fig.7.14a) to 2km in the east (Fig.7.14b). These imbricates have been greatly modified by the backfolding phase. One major anticline–syncline pair is shown in section A–A' (Fig.7.14a). This divides the section into three dip domains, one zone of overturned to subvertical bedding flanked by zones of gentle southerly dip where the beds are the right–way–up. In section B–B' (Fig.7.14b) however, the anticline hinge bifurcates. These folds trend E–W to ENE–WSW and plunge gently westward. Characteristically, the final stages of backfolding led to hinge accommodation difficulties which were resolved by breach or out–of–syncline thrusting in the anticlinal hinge zones. This has occurred at each hinge mapped. Hence, no flat–lying anticline hinges are observed. Instead one finds abrupt younging and dip confrontations.

At roadside outcrops near Toydabad (Fig.7.14a), several parasitic folds are observed. In the Lower Tertiary units to the south of Toydabad these folds verge northwards, whilst in the steeply dipping Kawagarh Formation rocks to the north of Toydabad the folds verge southwards. These parasites constrain the breached anticlinal backfold axis.

The third phase of deformation is related to NE–SW directed compression rather than the SSE–directed transport of the Hazara thrust wedge. This late compression relates to the propagation of the SW–vergent Kashmiri thrust wedge. Equivalent SW–vergent structures along strike to the southeast are emergent, e.g. the Kotli Thrust. However, to the northwest these Pliocene–Pleistocene thrusts have encountered the older Hazara thrust wedge which was emplaced during the Upper Miocene. Furthermore, due to the backfolding event the Hazara thrust wedge is characterised by subvertical rather than subhorizontal bedding orientations which strike parallel to the NE–SW compression direction.

Such a template is difficult to imbricate and the result is the corrugation of the region by a set of cross–folds. The plunge of these small scale crumples depends upon whether

one is situated on the steeply-dipping overturned or gently-dipping right-way-up backfold limb. On the former these crossfolds plunge steeply to the south-southeast. On the latter they plunge gently towards the south-southeast. At the backfold axis the cross-folds pass through horizontal or north-dipping plunges.

The cross-folds are excellently exposed on bedding plane exposures just south of Dunga Gali on the old Dunga Gali-Khanspur road. They are also well-developed to the north-east of Kundla, on the Kundla-Dunga Gali road. They occur on a wavelength of 200-1000m and are symmetrical. Overall, the cross-folds represent a NE-SW shortening of 12-15%.

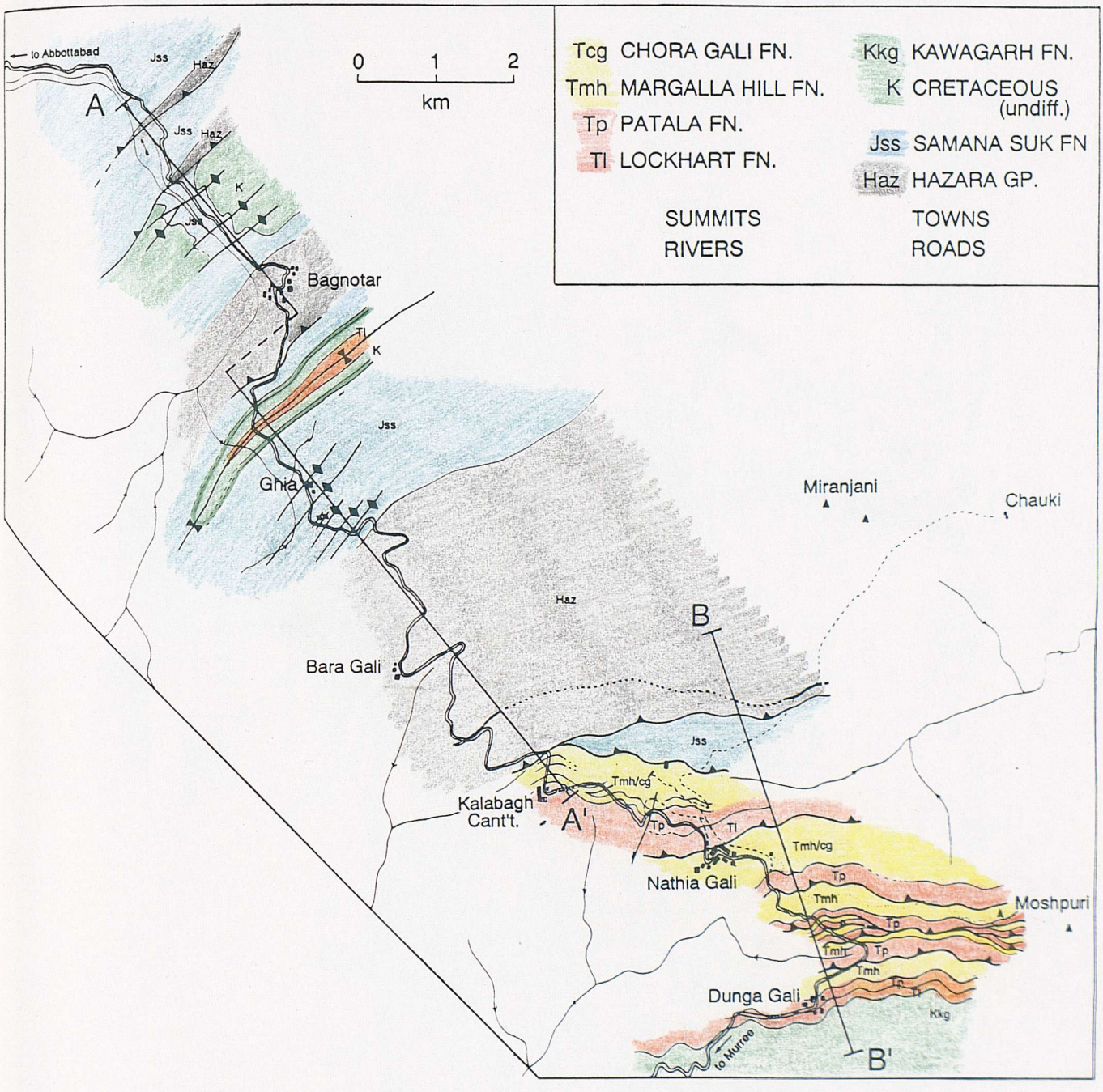
7.1.5. Dunga Gali - Bagnotar : the Slate Zone.

The area described in this section is the most northerly and therefore most internal part of the study area (Fig.7.15). It can be considered in two parts, the footwall and hangingwall of the Nathia Gali Thrust (NGT). Detailed mapping has been carried out in the NGT footwall zone where the lower Tertiary succession was imbricated during movement of the overriding NGT slab. A reconnaissance survey of the NGT hangingwall was carried out to establish the large scale structural relationships between the "Slate" and "Nummulitic" Zones of Middlemiss (1896).

The NGT actually outcrops 2km to the northwest of Nathia Gali village (Fig.7.15). It is exposed on the Nathia Gali-Abbottabad road just the north of Kalabagh where it juxtaposes Eocambrian Hazara Formation greywackes over Eocene Nummulitic shaley lime-stones. It can be traced from here east-northeastwards to the Nathia Gali-Chauki road where it juxtaposes Hazara Formation over Jurassic Samana Suk Formation. Between these localities a footwall splay comprising the Hazara and Samana Suk Formations branches with the NGT.

The introduction of the Hazara Formation indicates that the MBT, or Hazara Detachment, has ramped from Jurassic to Eocambrian horizons towards the north. Since the Jurassic and Eocambrian sequences are in direct unconformable contact the minimum height of

Fig.7.15. MAP OF THE SLATE ZONE



Sections A–A' and B–B' are shown in Figures 7.16a&b respectively.

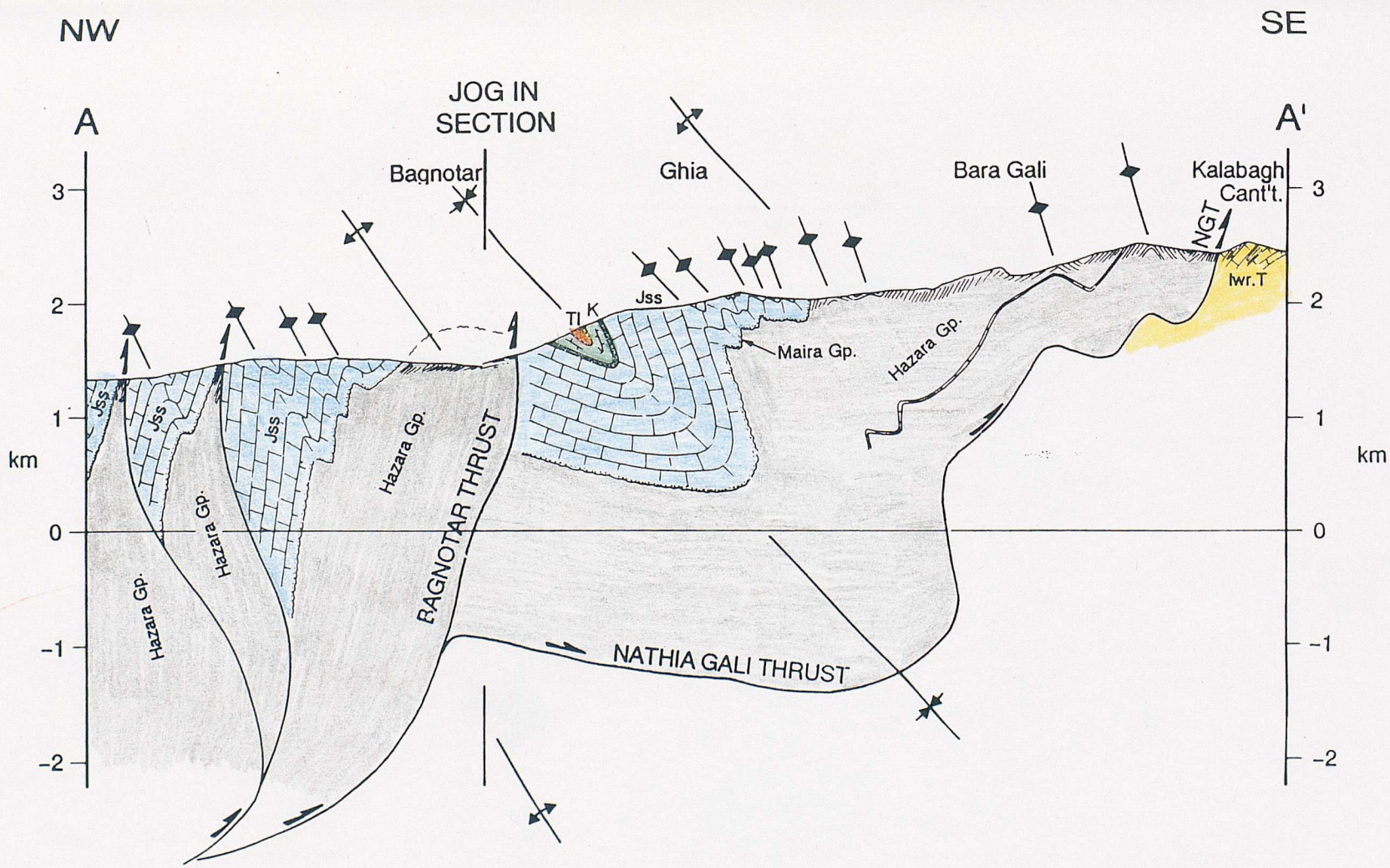


Fig.7.16a. SECTION A–A' ACROSS THE SLATE ZONE FROM BAGNOTAR TO KALABAGH CANTONMENT

For location see Fig.7.15.

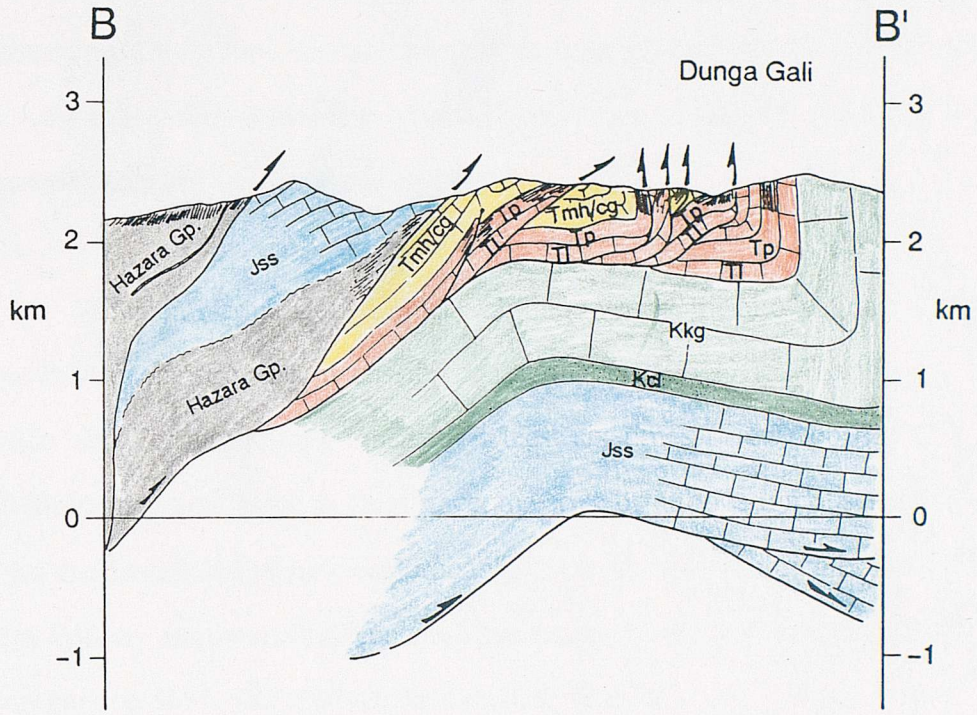


Fig.7.16b. SECTION B–B' FROM NATHIA GALI TO DUNGA GALI

For location see Fig.7.15.

this frontal ramp is equivalent to the thickness of Hazara Formation exposed at the surface, i.e. approximately 1500m.

The unconformity between the Eocambrian and Jurassic successions is exposed 2km north of Bara Gali where the vertical quartzitic sandstones of the Maira Formation are in stratigraphic contact with the Hazara Formation slatey greywackes. The unconformity omits the Cambrian Abbottabad Group that is developed further to the north and the Permo–Triassic units that form a thick sequence towards the southwest.

Between Nathia Gali, Dunga Gali and the summit of Moshpuri numerous repetitions of the Palaeocene–Eocene succession can be mapped (Fig.7.15). At the surface the Patala and Margalla Hill Formations are repeated at least six times (Fig.7.16b). Although the Patala Formation forms the shear plane at the surface, the style of imbrication observed at Changla Gali suggests that the Lockhart Formation is probably involved in the imbricates at depth and that the entire lower Tertiary sequence has been detached from the underlying thick Cretaceous–Jurassic carbonate sequence. It is possible that the Lockhart Formation has deformed with the Mesozoic succession rather than lower Tertiary at this locality. The style of imbrication suggests piggy–back footwall collapse in front of the advancing NGT. The complexity of the surface structure is in contrast with the interpretation of the structure at depth where the Mesozoic units have undergone SE–vergent thrusting and subsequent backfolding at a much larger scale dictated by the thickness of the Mesozoic limestone units.

The late crossfolding event is well illustrated on the map of the Slate Zone (Fig.7.15). The steep bedding orientations of the imbricated strata led to the corrugation of the thrust pile rather than its re–imbrication during NE–SW compression. The corrugation is expressed as a fanning of bedding strike about an E–W trend. This deformation phase is best observed in Ayubia and is described in section 7.1.4.

The region between Nathia Gali and Ghia on the Nathia Gali–Abbottabad road is

characterised by more gentle bedding dips reflecting the true dip of bedding at depth. Structurally, this region lies on the right-way-up backfold limb between an antiformal axis at Ghia in the north and a synclinal axis at Dunga Gali in the south. A second set of smaller scale folds with subhorizontal to gently-plunging axes is related to the early SE-vergent thrusting phase. The NGT therefore has a folded but generally gentle northwestward-dipping trajectory until it is folded about the antiformal axis at Ghia. This axis is characterised by intense small scale folding of the Jurassic limestones exposed at the surface (Fig.7.17a). These parasitic fold axes plunge at 20-30°NE and have axial planes that dip at 55°SE on average. The folds change from z-type to s-type folds as one traverses the fold axis from northwest to southeast. A cleavage has developed during folding which refracts markedly between the interbedded limestones and marly shales of the Samana Suk Formation.

To the southeast of Bagnotar a syncline is exposed in the Jurassic-Palaeocene sequence. Its axis trends NE-SW and is subhorizontal. The axial plane dips at 40-45°SE. The fold is interpreted as a NW-vergent synclinal backfold. A much reduced 150m thick Cretaceous succession is exposed in the fold core. The upper Cretaceous strata are thin and overlain by the 1-2m thick pisolitic laterite of the Hangu Formation. Such rapid stratigraphic variations demand either syn-sedimentary extensional faulting of upper Cretaceous age or major compressional shortening in Himalayan times. The latter model would require major displacement on the NGT, but since it dies out into a tipfold 20km along strike (Latif, 1970; Izatt, 1990) it cannot account for significant shortening. Similarly no major thrusts are present in the vicinity of Changla Gali to account for the upper Cretaceous thickness variations of that area. It seems probable therefore that this variation in thickness was caused by synsedimentary extensional faulting despite the fact that no such faults were discovered in the field.

The Bagnotar Thrust (Fig.7.16a) truncates the Mesozoic strata in its footwall at a high angle and is interpreted as an out-of-sequence thrust that breached an antiformal backfold axis. The Bagnotar Thrust has a displacement of *c.* 1km which uplifts the Hazara Formation strata in its hangingwall to the surface.

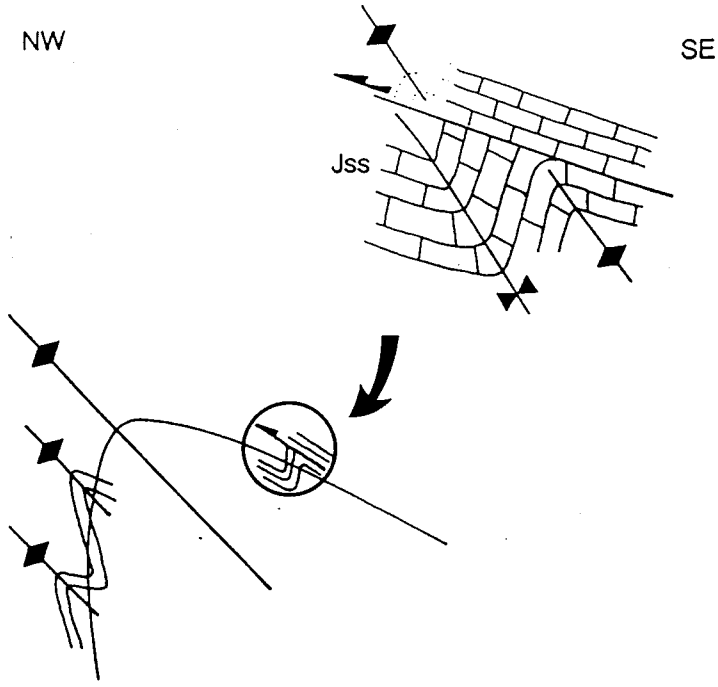


Fig.7.17a. PARASITIC BACKFOLDS VERGE TOWARDS INFERRED MAJOR AXIS, NEAR GHIA

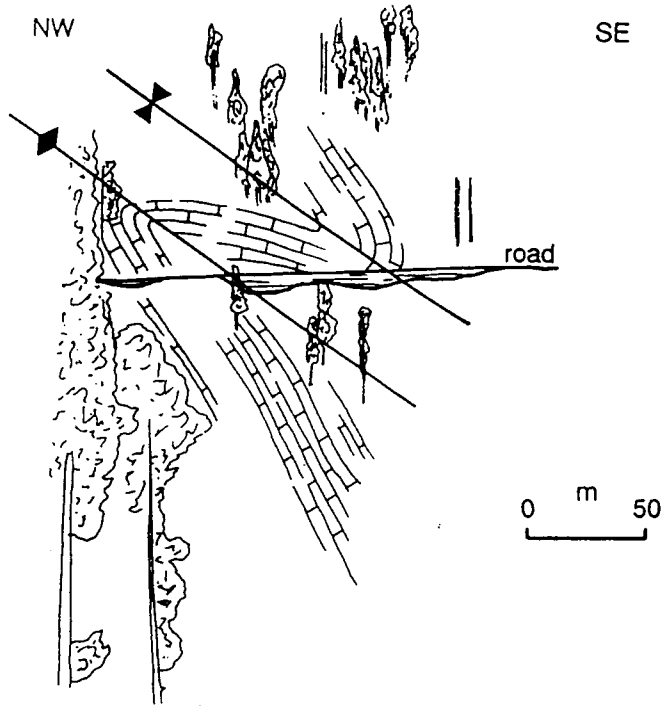


Fig.7.17b. SKETCH OF BACKFOLDS BETWEEN TOYDABAD AND KUNDLA

Northwest of Bagnotar one descends into the gorge of the Dor River – the incision of the Recent alluvial deposits testifying to recent uplift in excess of 200m. In this gorge the Jurassic strata are excellently exposed. They display magnificent chevron folds with axes subparallel to the major backfold axes (Fig.7.16a). The general orientation of the bedding is subvertical but the folding has generated panels of gently SE–dipping strata. On the road section small scale hinge accommodation structures are well–exposed.

The Slate Zone is therefore characterised by large scale SE–vergent thrusting and folding. Evidence of NE–SW compression is also observed. In the Jurassic exposures of the NGT nappe subhorizontal, WNW–trending fold axes are present. On a larger scale, a major NW–SE trending anticlinal fold is inferred from the opposing plunge of the NE–SW trending fold axes in the Dor River gorge section. It is clear therefore that both cross–folding due to the advance of the Kashmiri thrust wedge and backfolding related to the propagation of the Hazara–Potwar sole thrust at the basement–cover detachment are both observed as far north as the Slate Zone.

7.1.6. Summary of the Galiat Region.

The Galiat region comprises a highly–deformed but only slightly– or non–metamorphosed pile of sediments, ranging in age from Late Precambrian to Early Tertiary. The region lies within the Lesser Himalayan Zone of Gansser (1964) and is therefore bounded to the south by the equivalent of the MBT. This structure is the sole thrust to a SSE–vergent system of thrusts and folds that were imbricated and then translated at least 80km across the molasse basin. Continued shortening was accommodated by collapse in the footwall of the MBT and the development of a duplex between a floor–thrust at the basement–cover unconformity and a roof–thrust close to the MBT. During the latter stages of duplex formation the roof–thrust became a NW–vergent "passive" backshear. This led to the backfolding of the Lesser Himalayan Zone in the hangingingwall of the duplex.

The MBT, despite accommodating major displacement, is not observed as an emergent forethrust in N. Pakistan. Instead, it has been backfolded such that its tip is preserved in the subsurface and the carbonate/molasse transition at the surface is an anticlinal culmination. The cores of such culminations have been eroded to expose the underlying imbricates and thus the MBT, or Murree Thrust, plane is locally exposed. The tip of the MBT, however, remains concealed at depth to the south. The entire imbricated thrust pile of the Galiat has been backfolded as far north as Bagnotar, the northern limit of this project area. The bedding throughout the Galiat is generally steeply-dipping and only becomes subhorizontal on the southern limbs of anticlinal backfolds. The anticlinal backfold axes are prone to breach faulting.

Cross-folding due to SW-vergent compression has affected the Galiat during Pliocene times. Due to the backfolded nature of the Galiat, the superimposed SW-NE directed compression is expressed as a gentle corrugation of the bedding planes accommodating *c.*12% shortening. These folds are generally steeply-plunging, but since this ph phase has been superimposed upon the earlier backfolding the crossfold axes trace out the bedding profile of the backfolds. The folding therefore approximates to Type 2 of Ramsay's (1962) classification.

This phase of late stage deformation is also expressed by wholesale uplift of the Galiat that resulted in the incision of a new drainage system and the exposure of the MBT along the western limb of the HKS.

To summarise :-

1) the surface exposures in the Galiat lie in the hangingwall of the MBT (Hazara Detachment) which is not an emergent thrust, but is backfolded to form culminations.

2) the Galiat suffered three deformation phases -

a) a SE-SSE vergent phase of imbrication in the hangingwall of the MBT,

b) a NW-vergent backfolding event and

c) a SW-vergent crossfolding event.

3) major thickness variations in the upper Cretaceous succession are probably related to synsedimentary extensional faulting.

7.2. STRUCTURAL DATA FROM THE NPDZ.

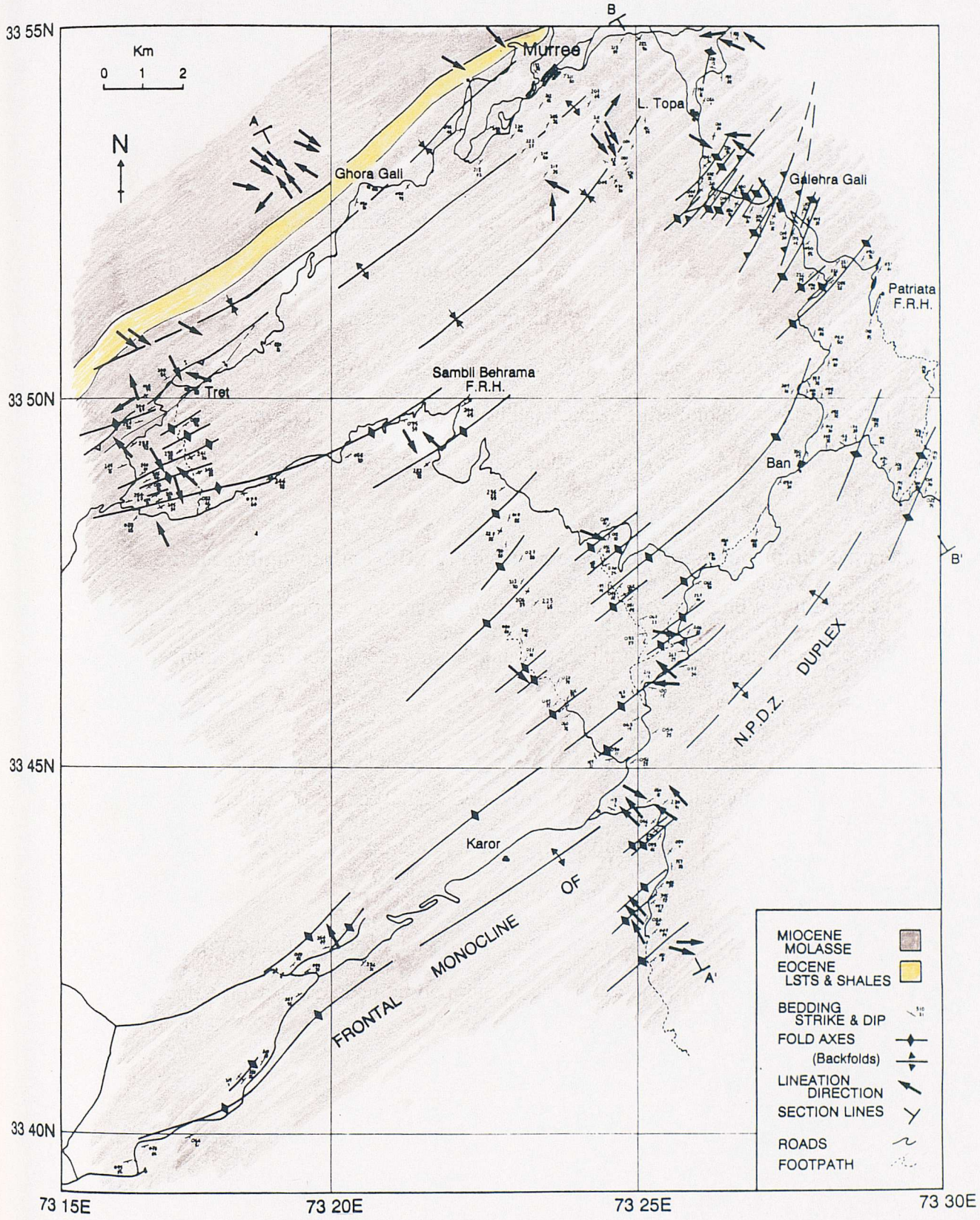
The Potwar Plateau has been studied in detail since the discovery of hydrocarbons. Exploration activity has, however, been centred on the western and central parts of the North Potwar Deformation Zone (NPDZ) and the eastern NPDZ has been largely ignored due to unpromising surface geology. Seismic surveys across the central NPDZ, completed during the mid-1970's, were made available to the author and were useful in constraining both the depth to basement and the imbricated nature of the Cambrian-Eocene carapace. Information from these profiles has been cautiously extrapolated to the area considered here.

The eastern part of the NPDZ which is considered here, is bounded to the east by the Jhelum river and to the west by the Islamabad Highway. The abrupt topographic break of the Margalla Hills marks its northern limit whilst in the south the NPDZ is bounded by the Soan syncline. During three field seasons in 1987, 1988/89 and 1990 a network of structural traverses across the eastern NPDZ was completed, covering both the flat-lying zone to the east of Islamabad and the uplifted foothills farther to the east.

Approximately 4km of Siwalik Group sediments have been eroded during the deformation of the NPDZ and throughout the entire region only Rawalpindi Group molasse is exposed. This is remarkable since the average bedding dip across the 20km wide NPDZ exceeds 45°. The molasse sequence has therefore been considerably thickened by imbrication and since Lower Tertiary sediments are nowhere exposed one can infer that the Rawalpindi Group has been decoupled from the underlying Cambrian-Eocene series. To account for the uplift of the NPDZ it is proposed that the Cambrian-Eocene sequence is imbricated to form a duplex at depth between two bounding detachments, one at the basement-cover contact and the other within the Rawalpindi Group molasse. In this model, the duplex roofthrust is a "passive" backthrust. The hangingwall molasse sediments suffered a strong NNW-vergent backshear whilst successive SSE-vergent Cambrian-Eocene imbricates were added to the duplex itself.

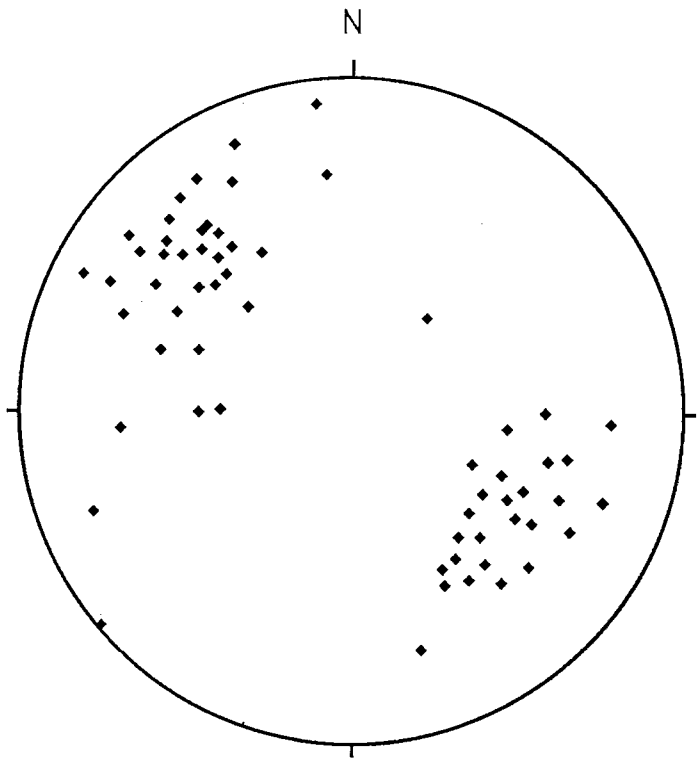
In the following pages the structure of the eastern NPDZ is described using three cross-

Fig.7.18. MAP OF THE UPLIFTED EASTERN N.P.D.Z., SOUTH OF MURREE.



Sections A–A' and B–B' are shown in Figures 7.19b&c respectively.

Mineral Lineation data from the eastern NPDZ



Key

- 62 Data from the North Potwar Deformation Zone

sections constructed parallel to the main tectonic transport direction (Figs.7.18 & 7.19a,b&c). Although NW–SE directed compression is responsible for majority of the structures present in the NPDZ, NE–SW compression has also affected this region resulting in the marked uplift of the easternmost cross–section described below. This recent uplift is manifest in the topography which rises from *c.*500m near Islamabad to *c.*2000m in the east.

7.2.1. The Bharakao–Thaliala Section.

Figure 7.19a is the westernmost of the three cross–sections presented here. Bharakao (Grid Ref. Q240620) is a bustling village lying on the Rawalpindi–Murree road. The northern part of the section is therefore easily accessible via minor roads and tracks leading from Bharakao towards Shahdara (G.R. Q230690) and Athai (G.R. 290610). The central part of the section is accessible from the Rawalpindi–Lehtrar road whilst access to the remainder is gained by taking the Thaliala (G.R Q270450) turn–off from the Islamabad Highway.

The southern limit of the NPDZ is clearly marked by the monoclinal flexure to the south of Thaliala. At this monocline the younging direction is consistently southwards but bedding dips fan from 70°SSE to 30°SSE in a distance of 1.5km, as one enters the core of the Soan syncline.

Between Thaliala and Kirpa (G.R. Q300480) a low topographic ridge occurs marking the surface outcrop of the Kamlial Formation. These green, flaggy, micaceous sands have not suffered the intense hydrafracking that is so ubiquitous in the overlying Siwaliks and are therefore quarried locally for building stone. This ridge marks the northern outcrop limit of the Siwalik Group sediments which have been eroded due to uplift associated with the deformation of the NPDZ.

At the southern limit of the NPDZ, immediately north of Kirpa, only NNW–vergent folds are observed. Towards the north, however, SSE–vergent folds and thrusts that have been refolded by NNW–vergent backfolds are observed. This change occurs near Thanda

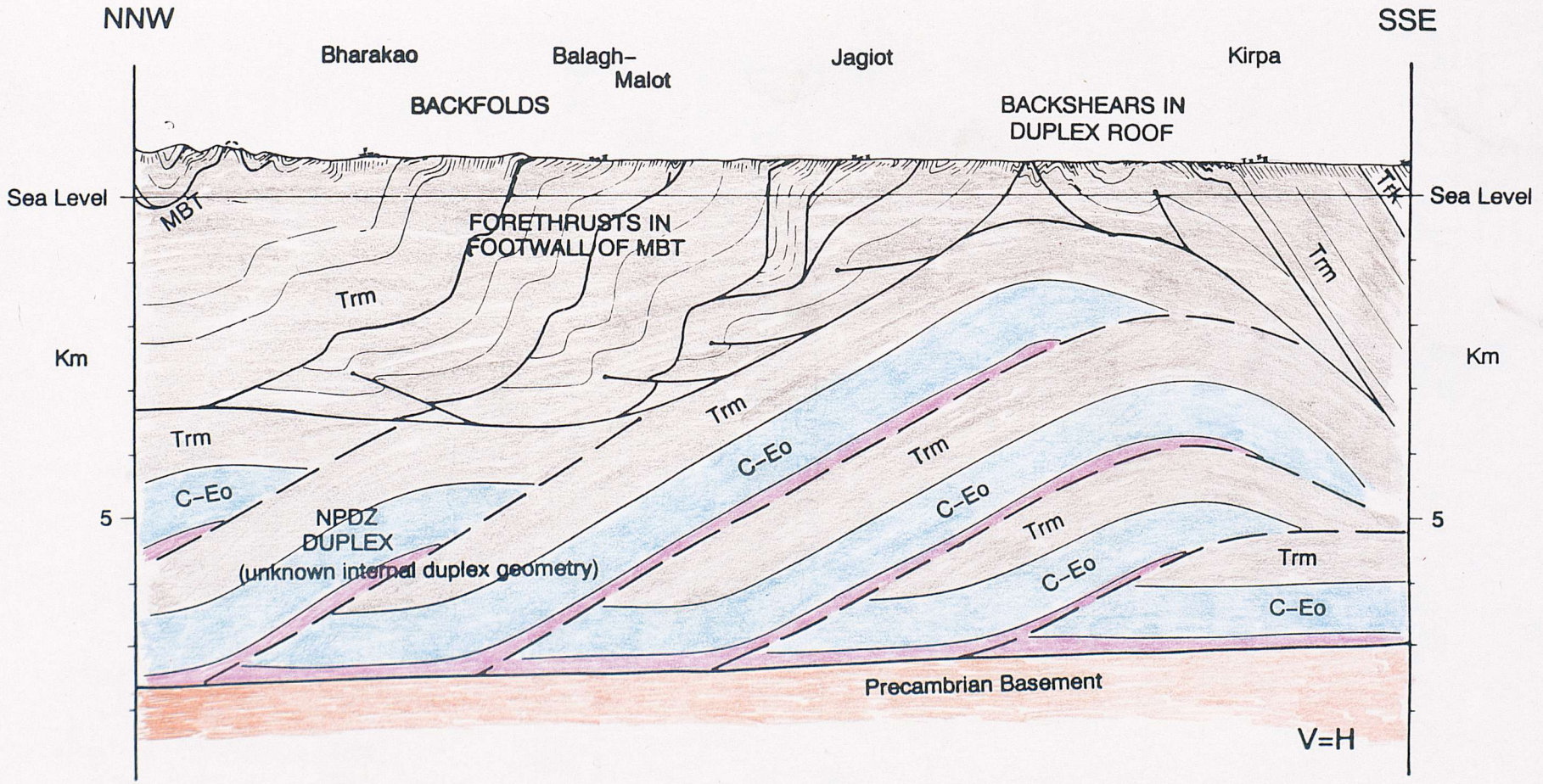


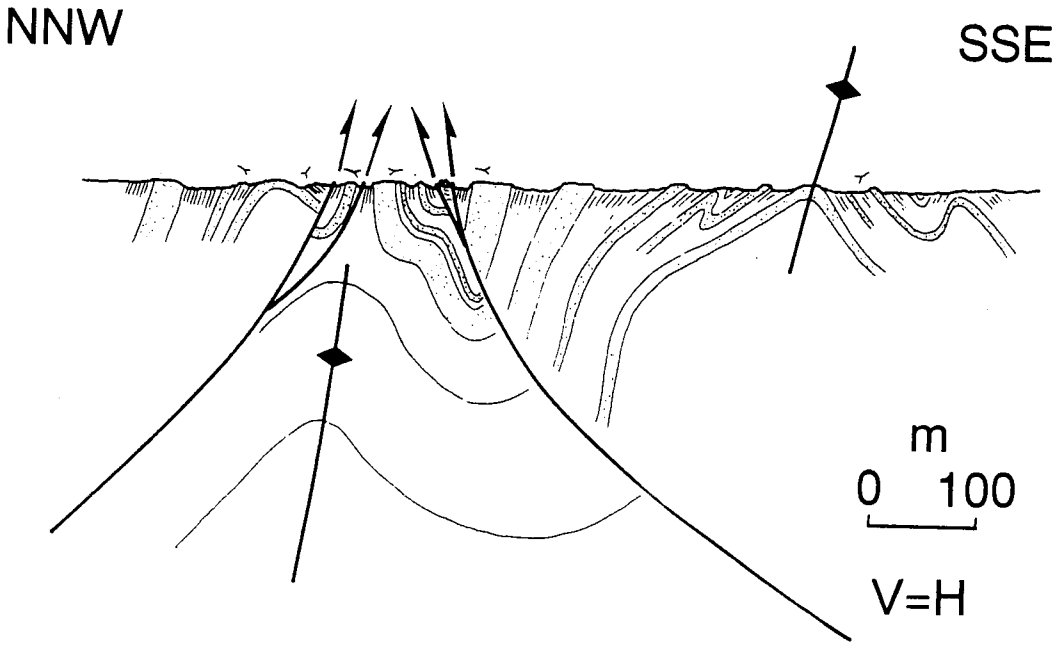
Fig.7.19a. SECTION ACROSS THE EASTERN N.P.D.Z. BETWEEN BHARAKAO AND KIRPA.

Trm Murree Formation
C–Eo Cambrian–Eocene

Fig.7.20. CARTOON AND DETAILED FIELD SKETCH ILLUSTRATING THE STRUCTURE AT THANDA PANI



Showing fold configuration
(1) before breaching



(2) after late breach thrusting

Pani (G.R. Q290520). This locality is inferred to mark the southern limit of SSE–vergent deformation related to the MBT zone. Structures related to this early deformation have been refolded by later NNW–vergent backfolds related to the propagation of the NPDZ duplex at depth.

This interference of deformation phases is neatly demonstrated by a pair of early 500m–wavelength, SSE–vergent anticlines located immediately north of the Rawalpindi–Lehtrar road at Thanda Pani (Fig.7.20). Both these structures possess parasitic folds that verge towards the anticlinal axes of the parent folds. However, during the later NNW–vergent deformation these folds have been disrupted. One anticline core is carried over an adjacent syncline core by a NNW–vergent breaching thrust whilst the other anticline core has been disrupted by steep reverse faults. The omission of fold axes throughout this zone results in the juxtaposition of opposite bedding younging directions. Strong shear lineations trending NNW–SSE (N159°E) were recorded on the thrust planes.

Further examples of the superposition of deformation phases are mapped to the north of this locality. In this part of the section bedding strikes ENE–WSW and dips steeply towards the NNW except for two narrow zones, near Jagiot (G.R. Q280550) and Balagh/Malot (G.R. Q260590), where shallower dips (25–35°) are recorded. Just north of Jagiot bedding is folded around NNW–dipping axes in the footwall of two SSE–vergent thrusts. Calcite fibres from each of the shear zones are oriented within the range N143°E–N183°E and N162°E–N189°E. Along strike to the WSW, near Kuri (G.R. Q250550), fibre lineations lie in the range N156°E–N175°E. The average lineation from 16 fibre outcrops, and thus a good indicator of thrust transport direction, is N162°E. In the hangingwall of these thrusts the Murree Formation beds change orientation suddenly from dips of 35°NNW to approximately subvertical as they are folded about a SSE–dipping backfold axial plane. It is ambiguous as to whether this flexure is a true buckle fold or simply a fault–bend fold above a thrust ramp.

A similar flexure is seen at the Balagh/Malot locality where bedding swings from dips of 25°NNW to 75°NNW. A single thrust is mapped at the surface and thus, this locality does not

suffer from the same structural complexity as Jagiot. The fact that this style of flexural folding is repeated favours the backfolding model, especially in the light of field data from the Galiat region. The superposition of several footwall thrust ramps could produce a similar outcrop pattern but the absence of the corresponding hangingwall ramp structures, perhaps due to erosion, detracts from this model.

At the extreme northern end of this section Eocene carbonates in the hangingwall of the MBT are depicted (Fig.7.19a). This is inferred from surface mapping immediately to the north of this section line where Eocene strata outcrop at the surface. At this locality the MBT appears to be a shallow, SSE-dipping, SSE-vergent fault. The fact that Eocene beds are downthrown to be juxtaposed beside Miocene molasse indicates that the fault is, in fact, a refolded thrust. Since Eocene strata do not outcrop anywhere to the south within the NPDZ, they must be truncated against the MBT forming a hangingwall ramp. In the footwall of the MBT only Miocene molasse sediments are involved in the thrust imbricates. The MBT zone was fully described in section 7.1.

It is not possible to constrain the internal geometry of the NPDZ duplex at depth without good quality seismic data. Hence the internal duplex geometry shown (Figs.7.19a,b&c) is purely schematic. One is, however, able to constrain the duplex roof geometry. The monoclinal northern limb of the Soan syncline reflects the shape of the duplex roof at depth. Approximately 2km of Murree Formation molasse is exposed at the surface and this, in conjunction with published commercial seismic data (Lillie et al, 1987; Baker, 1987), is used to predict the depth to detachment. Towards the north this level becomes more speculative since the sediments in the hangingwall of this detachment were deformed during the emplacement of the MBT.

The shortening accommodated by the duplex can be estimated from the surface deformation since this is accommodated at the surface by the backfolds. A major emergent backthrust is not observed at the frontal triangle zone of the duplex. The majority of the shortening within the duplex must be balanced by NNW-vergent backfolding and reactivated SSE-vergent thrusts at

the surface. In addition, estimation of the amount of material eroded during duplex formation allows coarse area balancing methods to be utilised. These assumptions suggest that the duplex must be shortened by approximately 30%, implying that the duplex imbricates are 2–3km thick. Since the Cambrian–Eocene sequence is only 800–1000m in thickness (S.M.Naeem, *pers. comm.*) in this area approximately 1–2km of Murree Formation molasse is inferred to be incorporated into the duplex itself. Although the depth to crystalline basement is constrained by both gravity data (Duroy, Farah & Lillie, 1989) and seismic data (Lillie et al, 1987), a further unknown is the thickness of Eocambrian sediments incorporated into the duplex.

The change in structural style between the northern and southern Potwar regions is attributed to an Eocambrian facies transition. In the NPDZ the Eocambrian may be absent and marked by an unconformity or may comprise the alternating sandstone and shale facies of the Hazara Slate Formation (Latif, 1973). Evaporites are, however, did not underlie the NPDZ during duplex formation. This poor constraint upon the Eocambrian strata could greatly the alter the internal duplex geometry.

7.2.2. The Danah–Lehtrar Section.

Although the basic structure illustrated in the Danah–Lehtrar section (Figs.7.19b & 7.18) is the same as that in the Bharakao–Kirpa section (Fig.7.19a) several important differences can be seen.

a) Firstly, **the topography of the Danah–Lehtrar section exceeds 1000m on average and includes ridges that exceed 1500m**. The average elevation of the Bharakao–Kirpa section is approximately 500m. This results from recent uplift generated by SW–vergent thrusting. In the line of section no stratal duplication is depicted, but major thickening is shown in the Kashmir Point–Nara section (Fig.7.17c) to the east.

b) Secondly, **numerous south–vergent folds are mapped in the MBT zone** in this section, whilst in the Bharakao–Kirpa section thrusting is dominant at the surface. This could be inter–

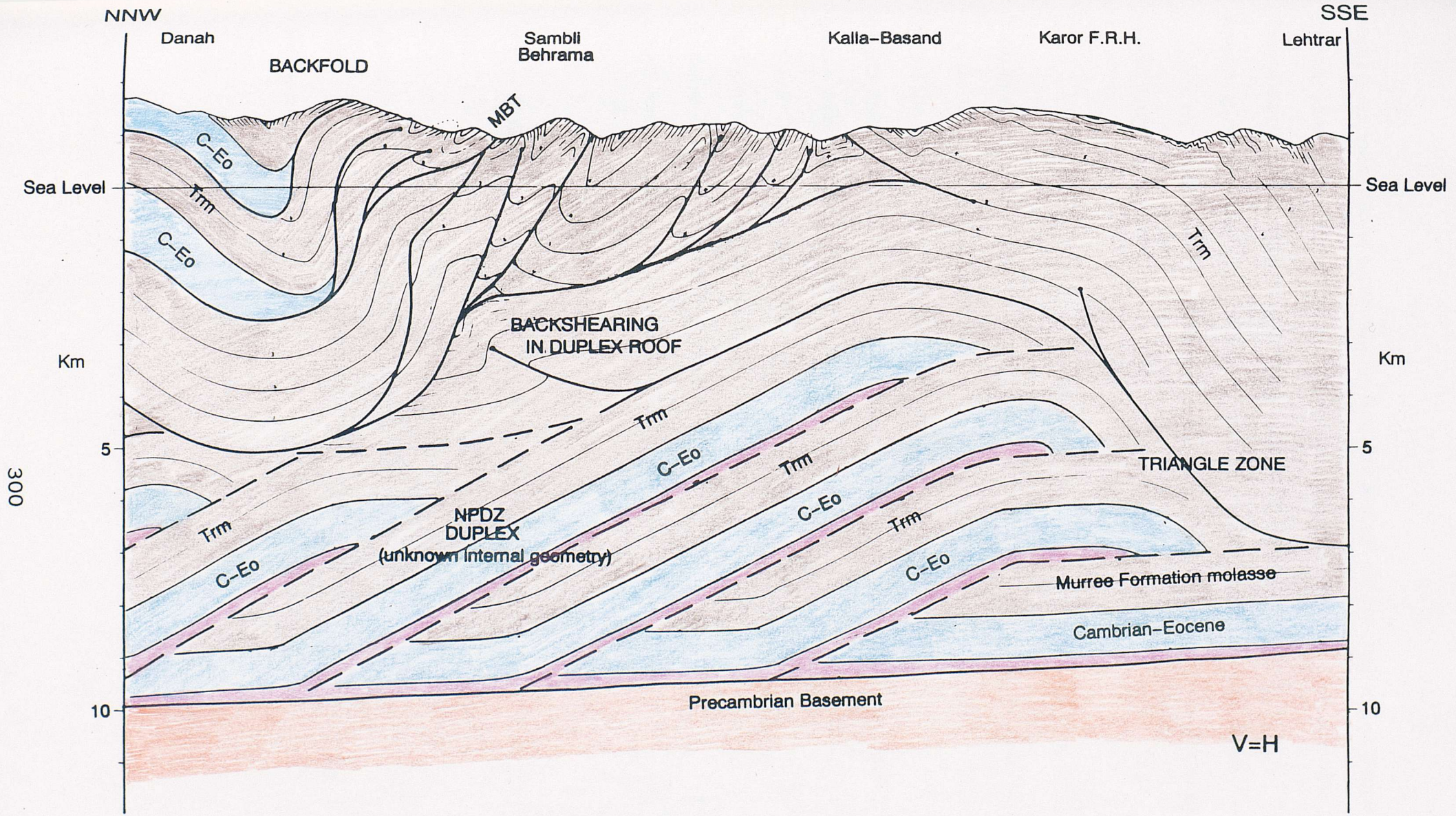


Fig.7.19b. SECTION ACROSS THE EASTERN N.P.D.Z. BETWEEN DANAH AND LEHTRAR.

preted as an indication of decreasing displacement towards the east with thrusts terminating in tipfolds. However, an important factor to consider is the level of erosion exposed in each section. The structures in both sections have been transported to the south–southeast by 20–25km due to shortening of the South Potwar region. Translation of wedge material on its dipping basal detachment will produce an increase in surface elevation. Erosion since the time of translation will expose deeper levels of the thrust wedge.

However, in the eastern Potwar Plateau, the interference of the SW–vergent Himalayan and SSE–vergent Pakistani thrust wedges means that the common basal detachment to these wedges does not dip solely to the north–northwest, but also towards the northeast. Across the eastern Potwar, the basement dip in a NNW–SSE plane is less than 1°, but it increases towards the north as the Indian crust responds to the load of the Pakistani thrust wedge. At any point the basement dip in a NE–SW plane is *c.*4–5°NE due to subsidence related to the load of the Kashmiri thrust wedge. This subsidence counteracted the uplift produced by NPDZ duplexing and minimised erosion in the east. Rather, the present relief of the easternmost NPDZ is related to Pleistocene uplift on the SW–vergent Kotli thrust system. This accounts for the deeper level of erosion in the Bharakao–Kirpa section compared with the Danah–Lehtrar section.

c) A further difference between the sections is the **depth to basement**. This has been mentioned above and is a result of the northeasterly dip of the basement–cover contact towards the Pir Panjal range of Kashmir. At the southern termination of the NPDZ duplex the basement lies at approximately 7km in Bharakao–Kirpa section (Fig.7.19a) and at 9km in the Danah–Lehtrar section (Fig.7.19b), implying a basement dip of 4–5°NE. This subsidence is reflected by molasse thickness variations.

7.2.3. The Kashmir Point–Nara Section.

This section (Figs.7.19c & 7.18) is the easternmost of the three sections constructed across the eastern NPDZ. Its surface geology also the most complex.

At the southern end of the section the predominant moderate–steep southerly dips reflect the S–dipping front of the NPDZ triangle zone at depth. The exact geometry of the triangle zone is disguised by backfolds such as the Nara anticline at the surface, that are interpreted as the tipfolds of thrust splays within the Murree Formation molasse. This structure is persistant for many kilometres along strike and probably represents an eroded Himalayan mountain front (Izatt, 1990).

The northern half of the section is much more complex than either the Bharakao–Kirpa (Fig.7.19a) or Danah–Lehtrar (Fig.7.19b) sections and the surface exposures are unlike anything observed to the west. The principal difference is the presence of overturned bedding between Lower Topa and Galehra Gali. These zones of overturned Murree Formation molasse lie in the footwall of S–vergent thrusts of the MBT zone. These thrusts clearly accommodate more shortening than their lateral equivalents to the west and their cut–off relationships indicate that they are late stage structures that breach pre–existing folds. Most of the beds are overturned towards the south. However, immediately south of Lower Topa, the strata are overturned towards the north. This is a result of late backfolding that post–dated the S–vergent breach thrusting event – these beds lying in the axial zone of SSE–dipping backfold axis. The degree of overturning (overturned bedding dips at 15°SSE) demands that another factor was at play and it is proposed that an early breaching backthrust must have existed but has since been eroded.

The SSE–vergent MBT wedge extended to the south of Galehra Gali. SSE–vergent folds at the surface in this area suggest the presence of a blind detachment at depth (Fig.7.19c).

The average elevation across this section is approximately 1500m, with peaks exceeding 2000m. The section has been recently uplifted by 1.5–2km. This has been accounted for

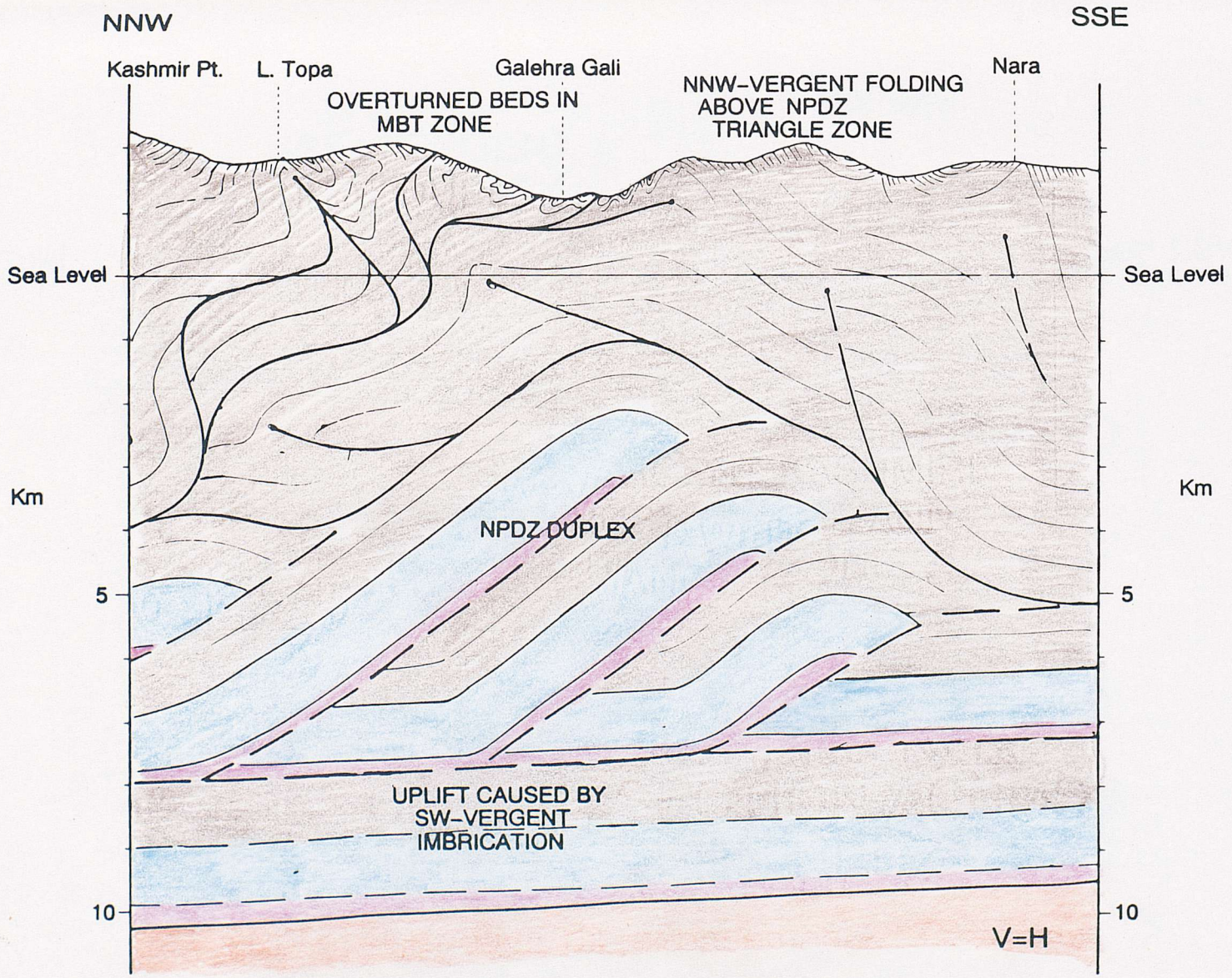


Fig.7.19c. SECTION ACROSS THE EASTERN N.P.D.Z. BETWEEN KASHMIR POINT AND NARA.

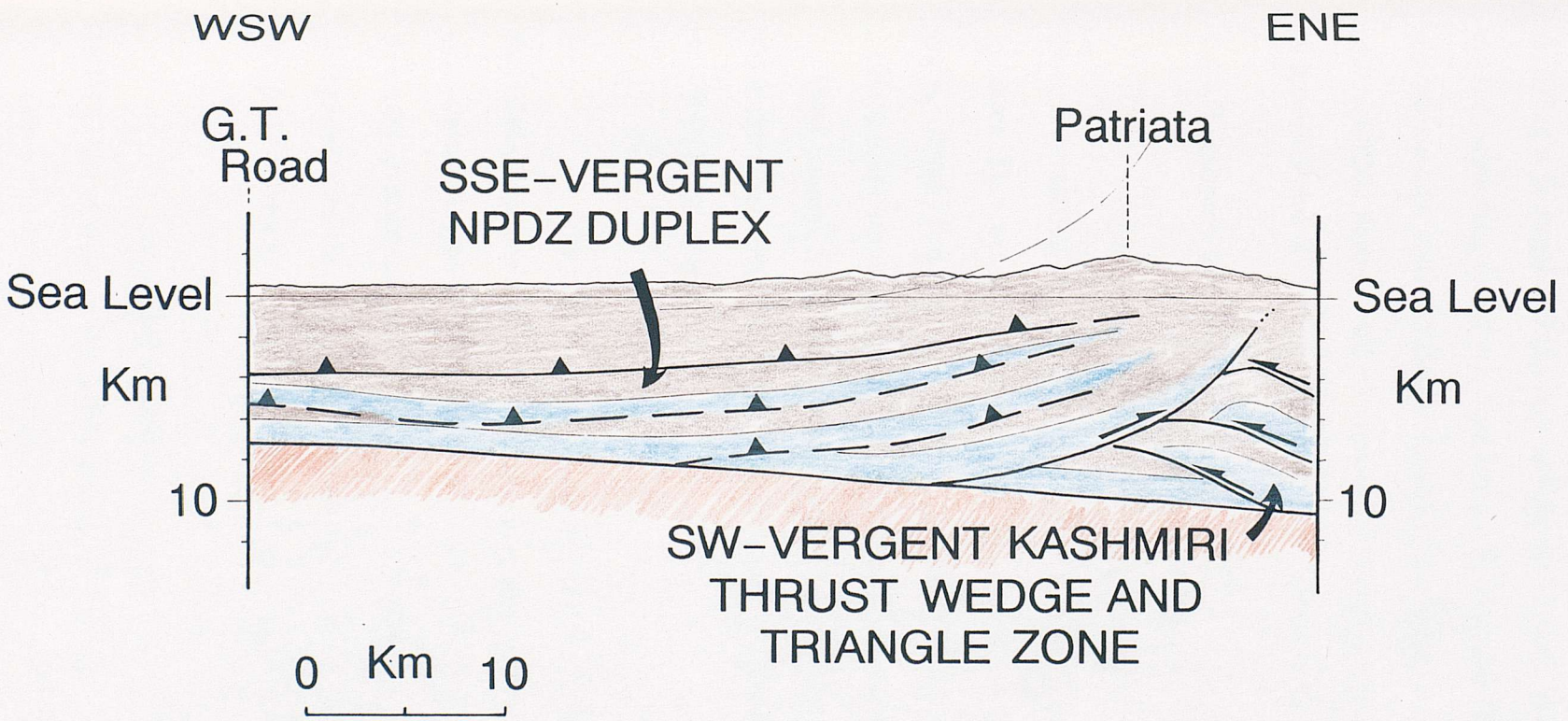


Fig.7.21. WSW–ENE TRENDING SKETCH SECTION ACROSS THE EASTERN NPDZ.

Note : 1) the NE dip of basement
2) the uplift of the easternmost NPDZ by the SW–vergent Kashmiri thrust wedge.

(Fig.7.21) by the introduction of material from out of the plane of section on SW-verging thrusts. Since the NPDZ duplex had been formed by the time of this Pleistocene SW-vergent thrusting it is suggested that the late thrusts formed an antiformal stack underneath the NPDZ duplex, producing the monoclinal western flank of the Hazara-Kashmir Syntaxis. In this area the monocline dips at only 20°SW, imparting this plunge to pre-existing WSW-ESE trending folds of the NPDZ. Farther east the monocline becomes subvertical as observed on the western flank of the HKS near Kohala, 24km north-northeast of Murree.

7.2.4. Transport Direction Indicators.

Transport direction indicators are difficult to find in the NPDZ because only the Murree Formation strata are exposed at the surface and therefore thrust surfaces are often inconspicuous. They are readily found in the vicinity of the Murree Thrust where intense shearing of the Kuldanna Formation and Chora Gali Formation shales is observed. Calcite fibres are the the most common shear lineation and occur in two settings, on flexural slip planes in fold limbs and on thrust planes. Since both these groups have developed in response to the same compressive stress the data points plot within the same field. There is a very strong alignment of these fibres in a NW-SE orientation.

The data comprises localities within the MBT zone and the NPDZ. The distribution of the lineation azimuths suggests a marked colinearity of the transport direction at the time of movement on the MBT and the deformation of the NPDZ (Fig.7.18). Although these zones are presently juxtaposed, after restoration the MBT and NPDZ regions lie at least 80km apart. Deformation of these zones spanned the period 10-5Ma. Thus, tectonic transport directions in the Hazara-Potwar thrust wedge varied little during this period.

Izatt (1990) noted the swing in lineation or transport direction from N210°E-N220°E in the western Kalachitta Ranges to N160°E in the Margalla Hills. The data presented here shows the continuation of this swing to the Murree Hills and eastern NPDZ where tectonic transport is towards the southeast (N130°E-N150°E) (Fig.7.22). Izatt (1990) argues that this

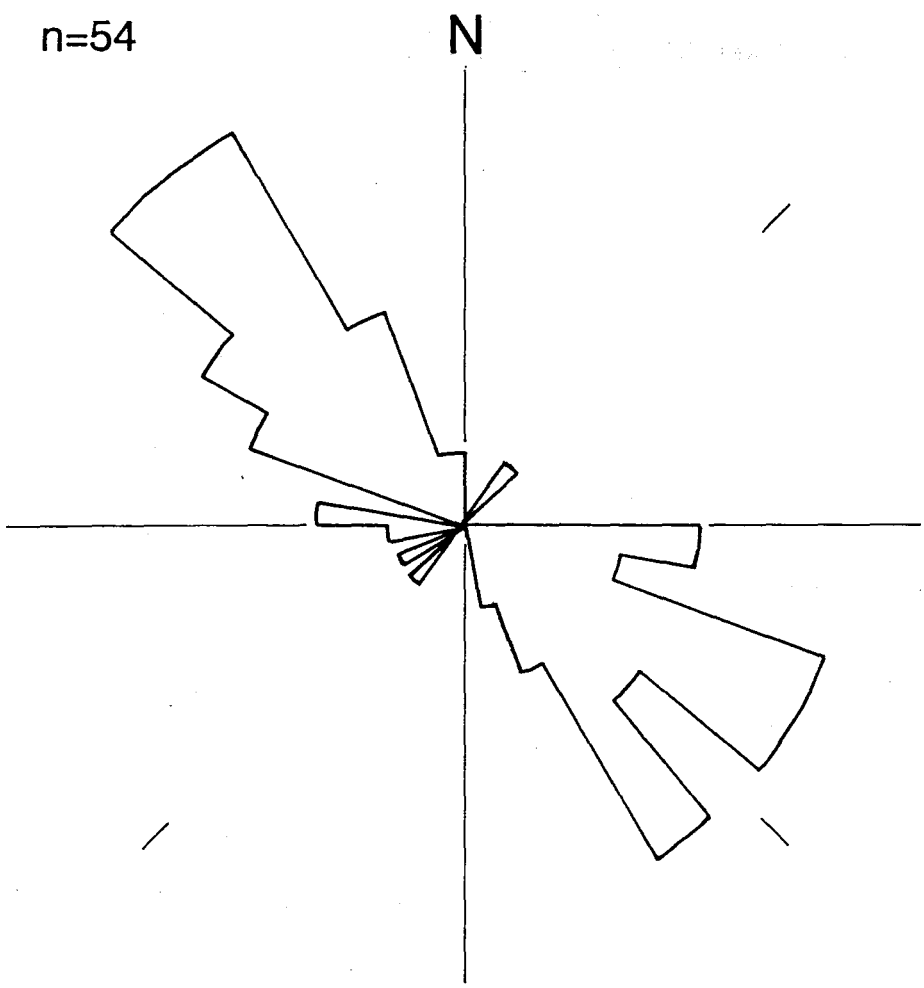


Fig.7.22. LINEATION DATA PLOT FOR THE EASTERN N.P.D.Z., indicating SE-vergent tectonic transport (N130°E-N150°E)

swing results from differential movement on the Potwar–Salt Range Detachment – displacement decreasing towards the east. This model is consistent with pinning of the SRT by the Pir Panjal thrust wedge to the northeast. The swing in transport direction reflects also the uplift of the eastern NPDZ and Murree Hills during the formation of the HKS. Refolding of a SSE–vergent lineation around a NW–SE axis would enhance the observed swing. This factor would not influence transport lineations in the western Potwar or Kalachitta Ranges.

7.2.5. Summary of the Eastern NPDZ Region.

1. The NPDZ formed prior to displacement on the SRT during the late Miocene (6–5Ma). The evidence for this is inferred rather than proven. The northern margin of the Soan syncline formed at 2Ma (Johnson, Raynolds & Burbank, 1986), the timing being tightly constrained palaeomagnetically, but this is thought to be a late out–of–sequence uplift rather than the date of main NPDZ imbrication.

2. The style of deformation in the NPDZ strongly suggests that it occurred prior to the exploitation of the Salt Range Formation as an evaporitic basal layer. The intense shortening observed at the surface is accommodated by SE–vergent imbricates at depth according to seismic interpretations. This is indicative of a strong basal resistance that argues that the basal detachment of the wedge lay within Hazara Group facies rather than Salt Range Formation facies rocks during NPDZ formation. Subsequent displacement on the Salt Range Thrust has carried the NPDZ duplex across the northern margin of the Eocambrian salt basin.

3. Since initial movement on the SRT is dated at 4.5Ma on the grounds of a palaeomagnetically–dated unconformity at the Baun syncline (Johnson, Raynolds & Burbank 1986), deformation of the NPDZ must have predated 4.5Ma.

4. Movement on the MBT was impossible after backfolding that is associated with NPDZ duplex formation. The timing of MBT displacement is poorly constrained. Izatt (1990) argues that the involvement of Chinji Formation strata in the Seri syncline on the southern margin of the Kalachitta Ranges indicates that the MBT is younger than 9–8Ma. It seems likely that the MBT

was active between *c.*8–6Ma and that the NPDZ deformed between *c.*6–4.5Ma.

5. The NPDZ is a zone of uplift (>3km) that caused the erosion of the entire Siwalik sequence. The southern margin of the duplex is marked by a frontal monocline that can be traced from Ban in the east, through Karor towards Kirpa in the west.

6. All the strata exposed at the surface in the eastern NPDZ lie in the hangingwall of the NPDZ duplex, in which the Cambrian–Eocene succession is imbricated. The depth to the duplex roof–thrust (which lies within the Murree Formation molasse) is unknown but the base of the molasse series becomes more deeply buried towards the east. This implies thickening of the molasse sequence towards the east which accords with models of foreland basin geometry based on gravity data (Duroy, Farah & Lillie, 1989). Towards the west, in the central Potwar, unpublished seismic profiles show that the top Eocene generally lies at *c.*2.3–2.6 sec TWT but becomes emergent as the Khaira Murat Thrust. Rather than reflecting variable uplift along strike, this may be a function of loading by the advancing Kashmiri thrust wedge and the corresponding thickness variation of the molasse cover from E to W across the basin.

7. The constraint of NPDZ duplexing to the period 6–4.5Ma means that the backfolding phase of deformation in the Galiat and the NPDZ is of the same age.

8. Since activation of the Salt Range Detachment at *c.*4.5Ma the NPDZ has been relatively quiescent. Palaeomagnetic dating of the northern limb of the Soan syncline at 2Ma (Johnson, Raynolds & Burbank, 1986) indicates that some late reactivation of the duplex has occurred. The low resistance to slip on the Salt Range Detachment means that little internal wedge thickening was required during wedge propagation since 4.5Ma.

7.3. STRUCTURAL DATA FROM THE SOUTHEASTERN POTWAR

This structure of this region, the most southerly part of the western limb of the Hazara–Kashmir Syntaxis (Fig.7.23), is well constrained by :–

1) surface structural mapping,

2) commercial reflection seismic data and

3) commercial well data.

Comparison between the surface and sub–surface data showed a marked contrast in structural style between deformation in the molasse and in the underlying shelf sequences. This occurs due to the development of a decollement surface close to the base of the Siwalik Group molasse succession. The sedimentary sequence underlying this intramolasse detachment comprises the Eocambrian salt, the Cambrian–Eocene predominantly "shelf" carapace and the Rawalpindi Group molasse. This sequence is detached from the Precambrian crystalline basement by the weak Eocambrian salt series and deforms in a more brittle style than the overlying Siwaliks.

Although a great variety of structures are observed on the seismic data, they appear to represent stages during the progressive evolution of a single structure. The structure initiates as a salt pillow that may nucleate above a basement offset (Wiltschko & Eastman, 1983; Lillie et al, 1987, Pennock et al, 1989). With continued halokinesis the pillow grows so that the overlying, brittle, Cambrian–Eocene carapace fractures to produce a pop–up. In the eastern Potwar the majority of the pop–up bounding thrusts either become bedding parallel at the intramolasse detachment, e.g.Ghungrila–Kallar, or terminate as box–shaped tipfolds. Occassionally they ramp through most of the Siwalik succession as a low angle thrust and become emergent, e.g. the Domeli Thrust.

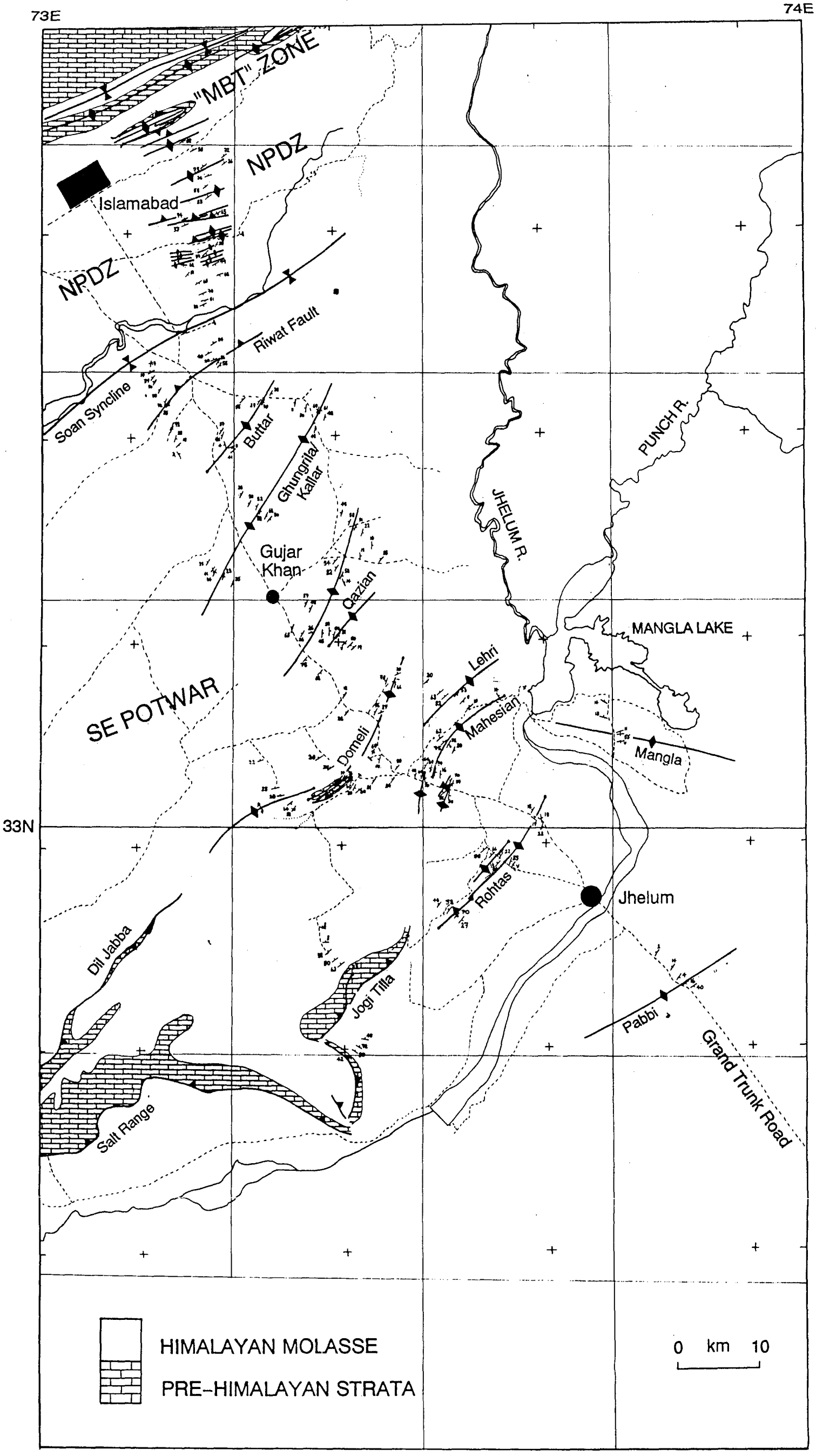


Fig.7.23. SURFACE DATA FROM THE EASTERN POTWAR.

Dashed lines represent roads

Hence, with increasing strain **three phases of structural evolution** are recognised :-

1) a salt pillow development,

2) amplification to form a pop-up and

3) breaching of the pop-up by a through-going thrust.

In the early stages of development uplift and vertical thickening are the dominant features of the structure; only in the final stage is significant horizontal shortening observed.

Eight major structures are recognised in the SE Potwar and each is described in the following sections.

7.3.1. Riwat.

The Riwat thrust was one of the earliest formed structures of the SE Potwar. Using magnetostratigraphic dating methods, Johnson et al (1986) indicate that the Riwat thrust was active during the period 3.4–2.7Ma. Since initial movement on the SRT and the formation of the Bhaun syncline has been dated at 4.5Ma (*op.cit.*), the Riwat structure developed as an out-of-sequence thrust behind the main Himalayan front.

Two traverses across the Riwat structure were made. Although the thrust plane itself was not observed in the field its location is tightly constrained by a change in younging direction (Fig.7.23). Steep bedding dips in the footwall of the thrust indicate that faulting occurred after an initial folding phase. In the hangingwall of the thrust, a 3.5km thick sequence of steeply-dipping molasse sediments is exposed before bedding dip gradually decreases towards the core of the Soan syncline. This shows that the Riwat thrust dips at >60°NW at the surface. Unpublished commercial seismic data shows the Riwat fault gradually flattening with depth (Fig.7.24). This recently acquired data suggests that the Cambrian–Eocene carbonate carapace is undisturbed under the Soan syncline. This demands that the Riwat thrust becomes bedding parallel within the Rawalpindi Group molasse sequence.

The evolution of the Riwat thrust is illustrated in Figure 7.25. The majority of

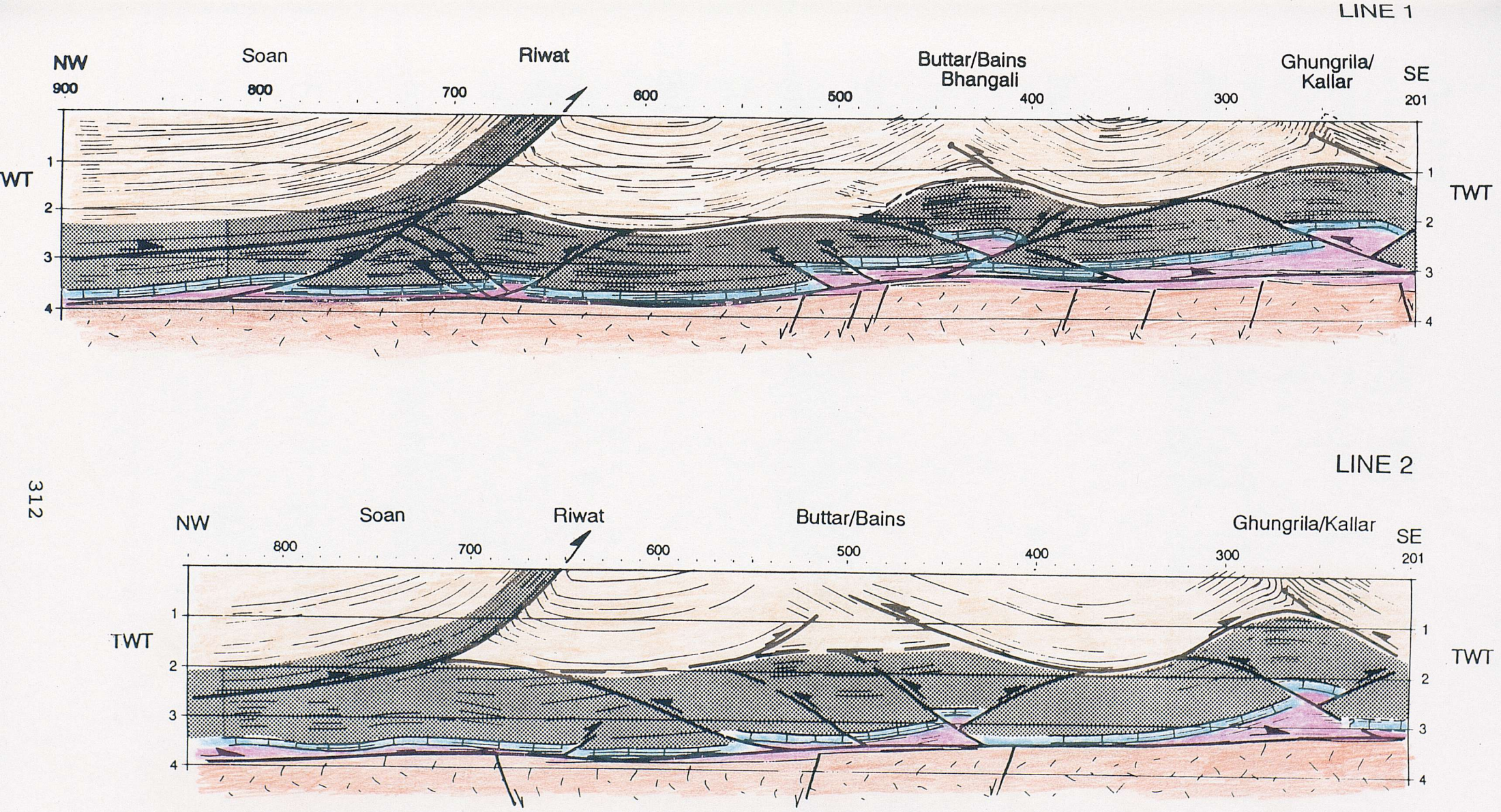


Fig.7.24. INTERPRETATIONS OF SEISMIC TIME SECTIONS 1 & 2, INCORPORATING SURFACE DATA.

(Ornamentation as before)

FIG. 7.34 (p.328)

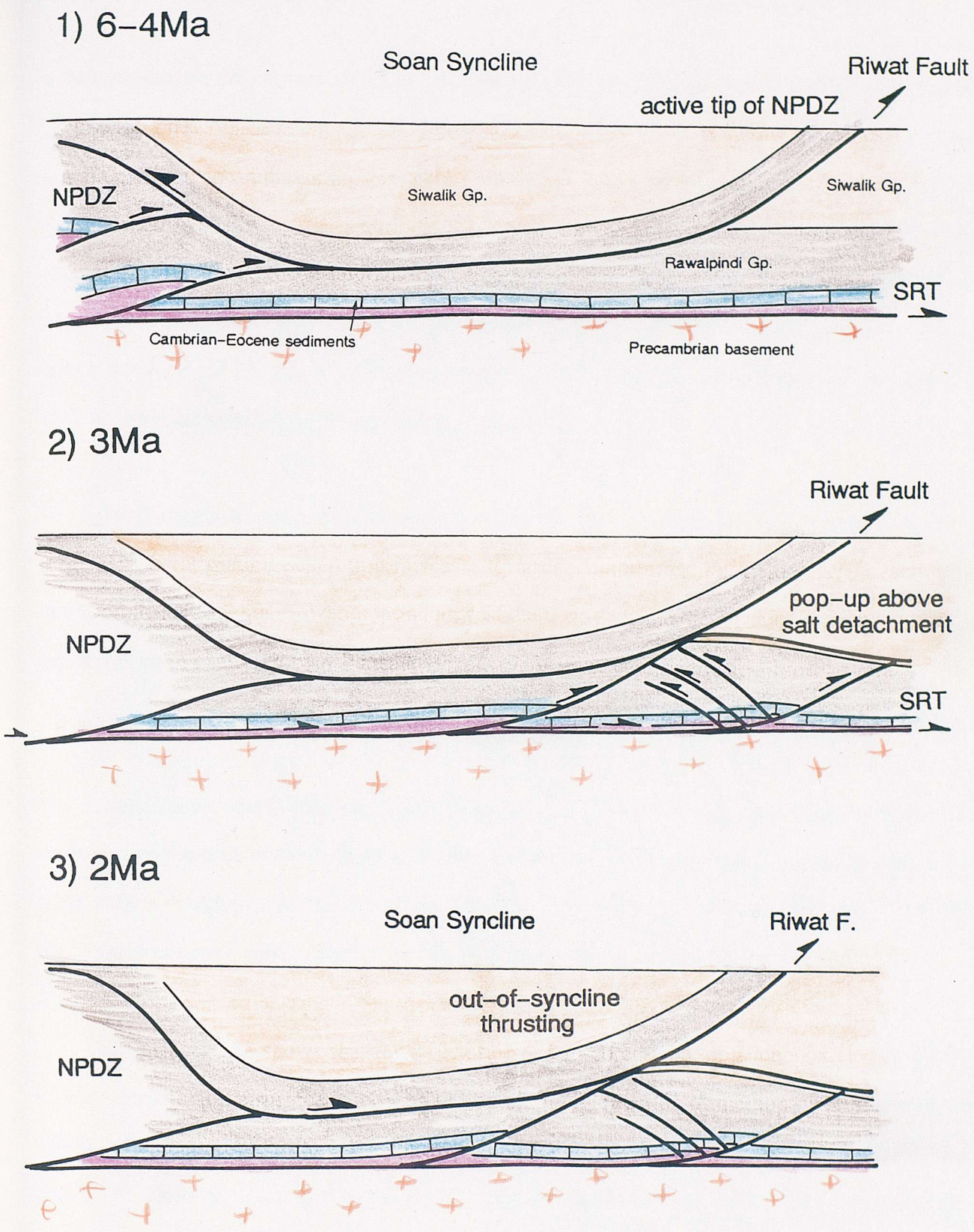


Fig.7.25. EVOLUTION OF THE RIWAT FAULT

1) Riwat F. is the emergent tip of NPDZ deformation which lies within the Rawalpindi Group molasse before ramping to the surface. SRT becomes active.
2) Deformation continues in the hangingwall of the SRT by pop–up formation beneath a basal Siwalik Group detachment. The upper part of the Riwat F. suffers minor reactivaton.
3) Northern limb of the Soan Syncline is tightened by NPDZ reactivation producing out–of–syncline displacement on Riwat F.

displacement on the Riwat thrust is associated with the last stages of NPDZ deformation (Fig.7.25, phase 1). Seismic Line 2 (Fig.7.24) shows that the Riwat fault becomes bedding-parallel within the molasse and probably links with the NPDZ triangle zone to the north. On Seismic Line 1 (Fig.7.24) an additional S-vergent thrust ramps from the basement-cover detachment, offsetting the Cambrian-Eocene sequence and reactivating the frontal part of the Riwat fault (Fig.7.25, phase 2). Pennock (1988) proposed an out-of-syncline origin for the Riwat thrust (Fig.7.25, phase 3). This may have occurred during a late tightening of the northern limb of the Soan syncline at 2Ma (Johnson et al, 1986).

Total displacement on the thrust exceeds 3km near the Grand Trunk Road section (Fig.7.23). Stratigraphic separation increases towards the northeast, but field access is denied as one approaches Kahuta. Towards the southwest the Soan syncline dies out and the Riwat thrust is not observed in central Potwar Plateau (Baker, 1988; Lillie et al, 1987).

7.3.2. Buttar.

Immediately south of Riwat lies the Buttar anticline. A single traverse of the structure was made just to the east of the G.T. road. At the surface the fold is a rounded, broad-hinged anticline with a roughly symmetrical geometry. The northwestern fold limb dips at 55-60°NW whilst the southeastern limb dips at 54°SE but flattens rapidly within 1km of the fold axis.

Subsurface data shows that the structure is a salt-cored pop-up (Pennock, 1988). The scale of the fold at the surface is too small to be simply a reflection of the pop-up culmination at depth. A small N-vergent thrust is therefore inferred to splay from the intramolasse detachment (Fig.7.24). This splay terminates at depth allowing the tipfold to diminish to the almost symmetrical structure observed in the field. There is considerable offset between the fold at the surface and the pop-up at depth in the northeast (Fig.7.24, line 2). This offset diminishes southwestwards (Fig.7.24, line 1). Such disharmony was described at Khaira Murat in the central Potwar by Izatt (1990).

Along strike to the southwest uplift on the pop-up diminishes, but shortening is accommodated by the oil-producing Bains-Bhangali structure.

7.3.3. Ghungrila-Kallar.

The Ghungrila-Kallar fold has been constrained by three surface traverses (Fig.7.23) and by two seismic profiles (Figs.7.24 & 7.26, lines 1 & 2).

In common with the Buttar and Qazian structures, **Ghungrila is a salt-cored pop-up whose bounding reverse faults** do not terminate at blind tips to produce a boxfold but, rather, **flatten to become bedding parallel detachments within the molasse succession**. This detachment is thought to lie close to the boundary between the Rawalpindi and Siwalik Groups within the basal Chinji Formation. Approximately 2km shortening (Fig.7.27) is accommodated by uplift of the fault-bounded triangle and doming of the overlying strata to form a broad culmination. From the seismic data one might predict a broad, flat-topped anticline at the surface. However, a zone of no seismic reflections at the anticlinal hinge corresponds to a region of steeply-dipping to overturned beds at the surface. This is characteristic of anticlinal hinge zones in the Potwar and is caused by small thrusts that splay off the intramolasse detachment horizon and terminate as tipfolds close to the surface. Such thrust splays can verge towards the foreland or hinterland and change sense along strike. This gives the fold a surface asymmetry that does not reflect the symmetrical form of the structure at depth.

Chinji Formation strata are exposed in the core of the fold. This unit is composed of dominant red shales and subordinate grey sands. Distinction between the L. Siwalik Formations is however difficult in the field.

Seismic data from 5km southwest of the G.T. road (Fig.7.24, line 2) shows the top of the Eocene carbonate sequence at 2sec TWT within the uplifted pop-up compared with the regional level of 3.2secs TWT. The carbonate carapace has therefore been uplifted by approximately 1.8-2.0km, assuming that the overlying molasse sediments have a seismic velocity of 3000-3300m/s

Fig.7.26. TWT CONTOUR MAP OF THE BASE MOLASSE, S.E. POTWAR

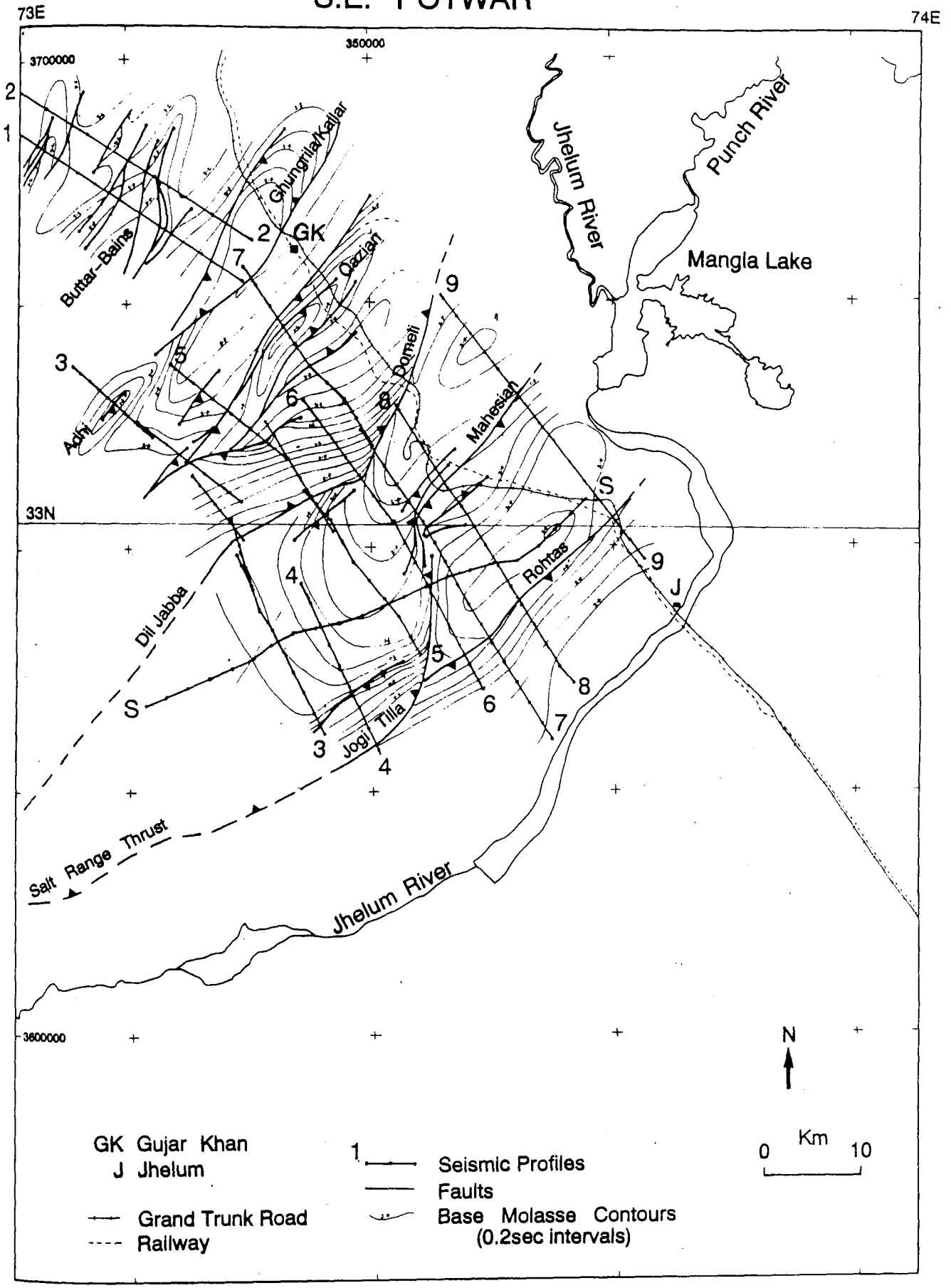
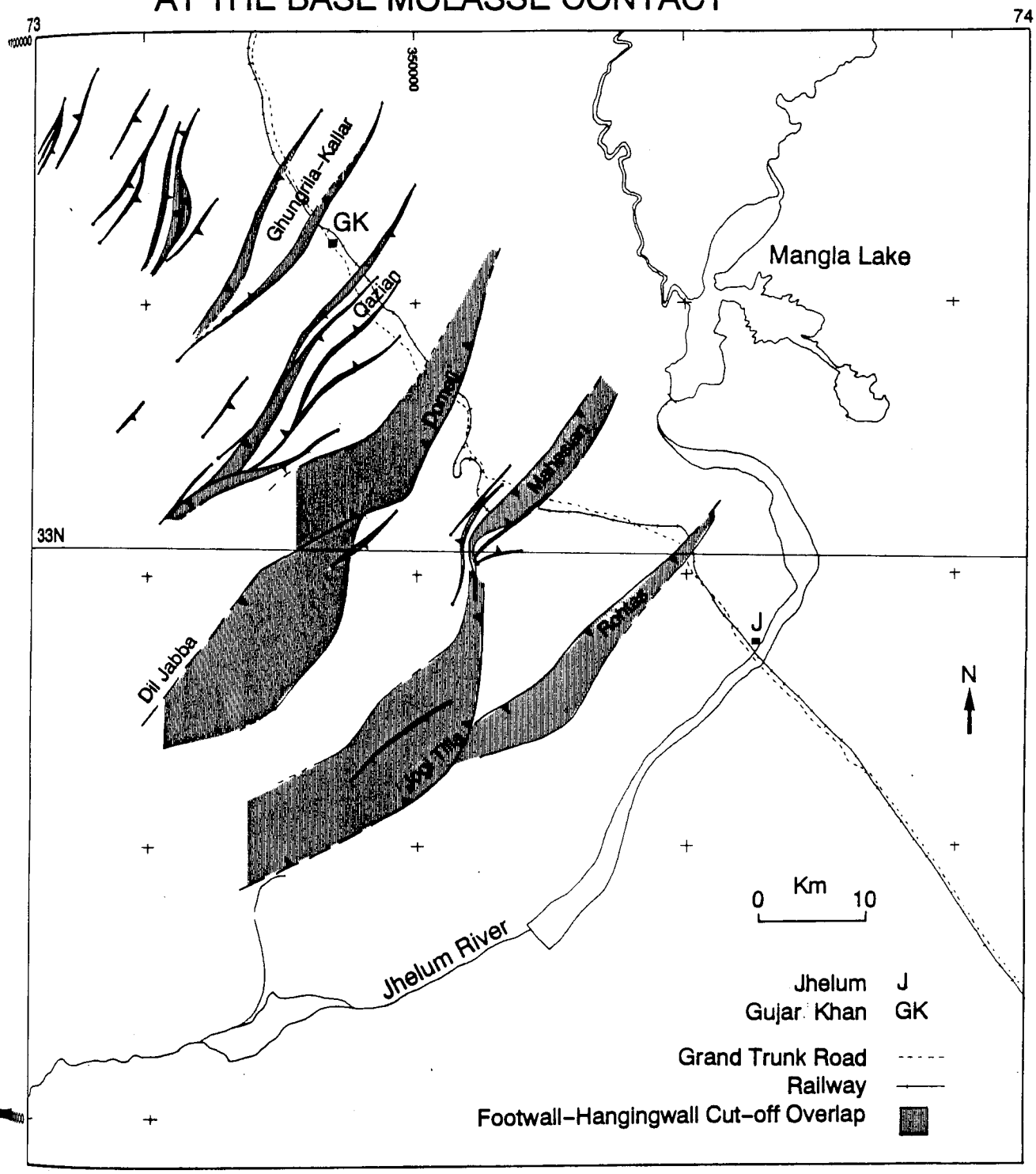


Fig.7.27. FAULT CUT-OFF MAP, S.E. POTWAR AT THE BASE MOLASSE CONTACT



(Pennock, 1987; Baker, 1987). According to well data, the Cambrian–Eocene sequence is approximately 500m thick in this region. The underlying Eocambrian Salt Range Formation varies considerably in thickness from negligible thickness under the synclines to over 2000m in the core of the Ghungrila pop–up. Crystalline basement lies at approximately 3.2–3.3secs TWT (equivalent to *c.*5.5km depth).

The regional basement dip towards the northeast is reflected by the plunge of the fold in that direction (Fig.7.26). Towards the southwest the uplift of the structure decreases temporarily before rising again at Adhi.

7.3.4. Qazian.

Four structural traverses were made across the Qazian structure, all of which lie on or to the east of the G.T. road (Fig.7.23). In addition, three seismic profiles cross the structure to the west of the G.T. road (Fig.7.26, Fig.7.28, line 3, Fig.7.29, line 5 and Fig.7.30, line 7). The Qazian structure is currently an active petroleum exploration target and, hence, access to subsurface data towards the east was not possible.

The geometry of the Qazian fold at the surface changes considerably along strike. Two thrusts splay from the flanks of a culmination in the intramolasse detachment and are marked by tipfolds at the surface. From field mapping of these folds, the NW–vergent splay becomes increasingly dominant along strike towards the northeast where the fold is strongly NW–vergent at the surface (Fig.7.31). At the G.T. road the SE–vergent splay is more dominant. Exposure of the crest of the Qazian culmination is poor in this area but shallowly dipping beds (20–25°) were mapped in the core between the two tipfolds. Bedding dips steeply to both northwest and southeast on the outer limbs of the Qazian culmination before flattening into the adjacent synclines exposing almost 3km of Siwalik stratigraphy from Chinji Formation to Dhok Pathan Formation.

The seismic data reveals the variation of the subsurface structure along strike. Towards the west, Qazian is simply a backthrust at depth draped by a monocline at the surface (Fig.7.28, line

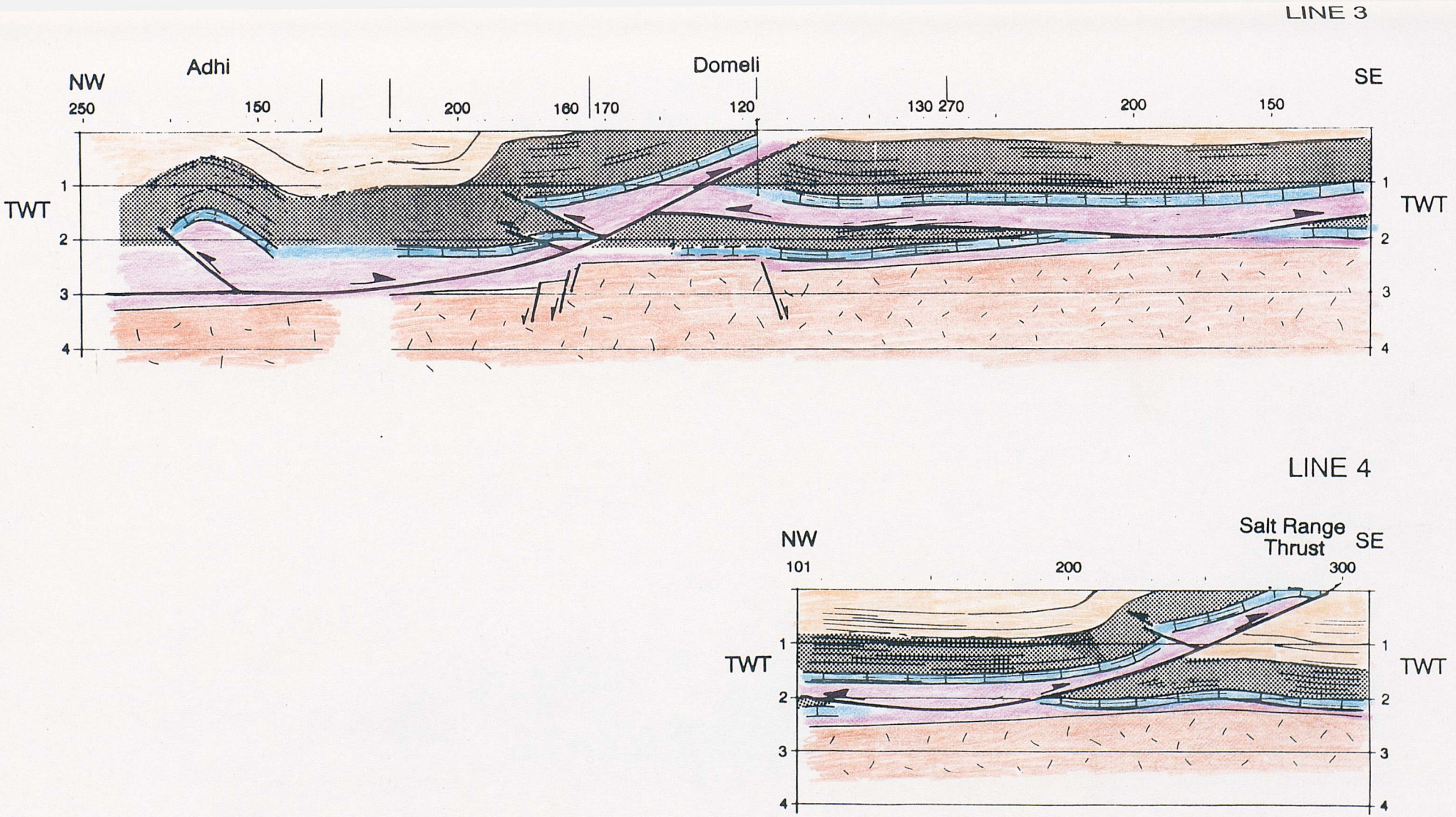


Fig.7.28. INTERPRETATIONS OF SEISMIC TIME SECTIONS 3 & 4, INCORPORATING SURFACE DATA.

(Ornamentation as before)

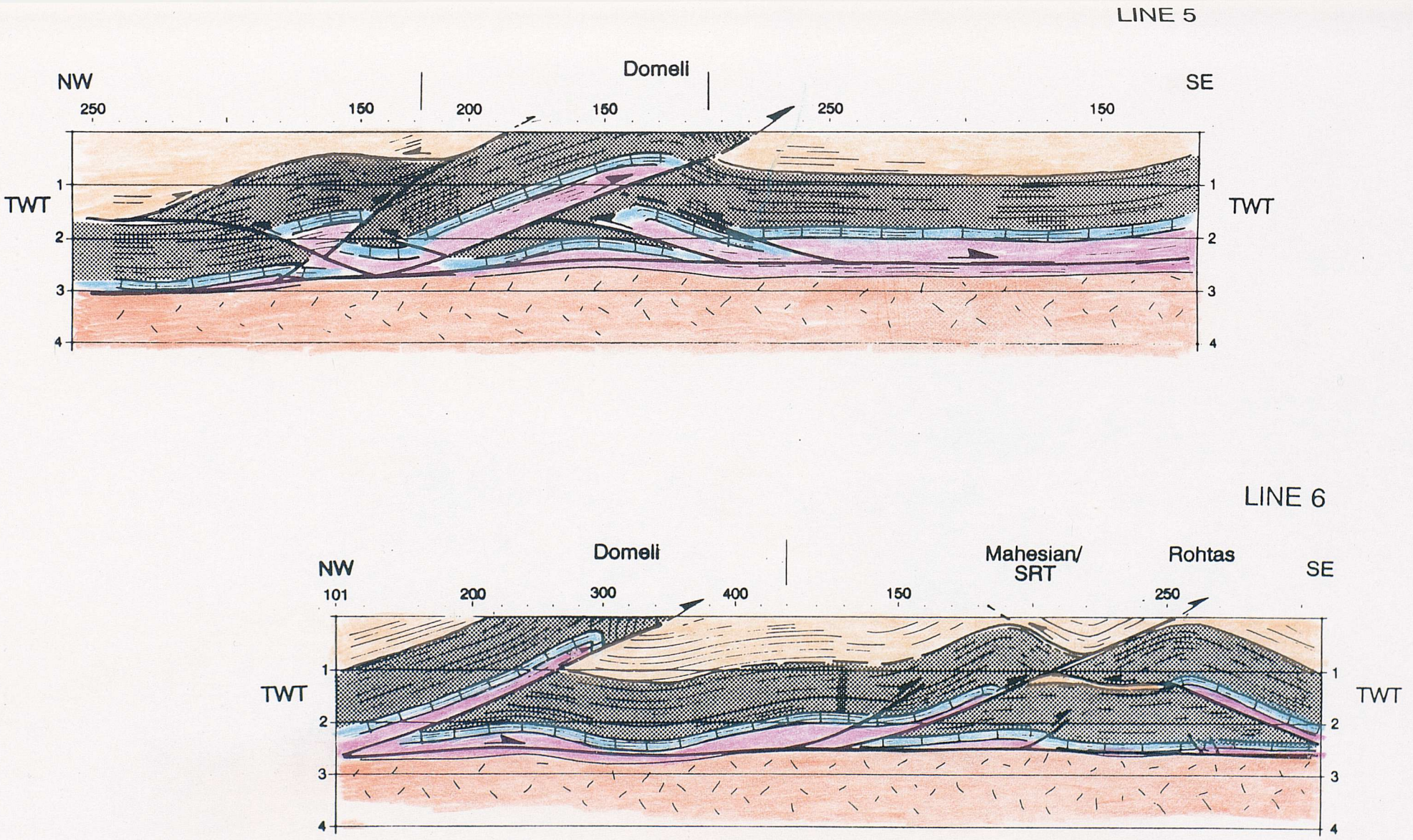
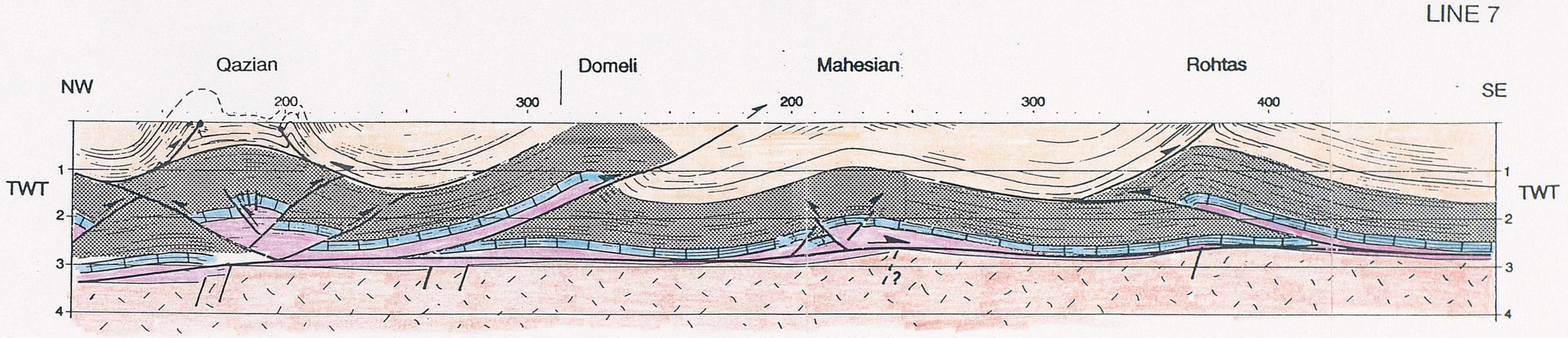


Fig.7.29. INTERPRETATIONS OF SEISMIC TIME SECTIONS 5 & 6, INCORPORATING SURFACE DATA.

(Ornamentation as before)



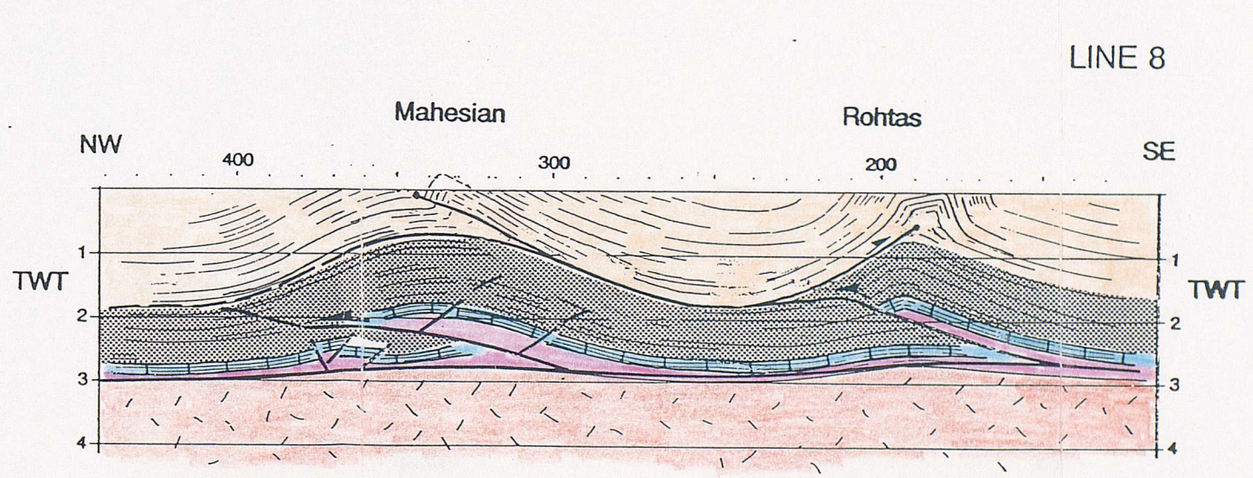
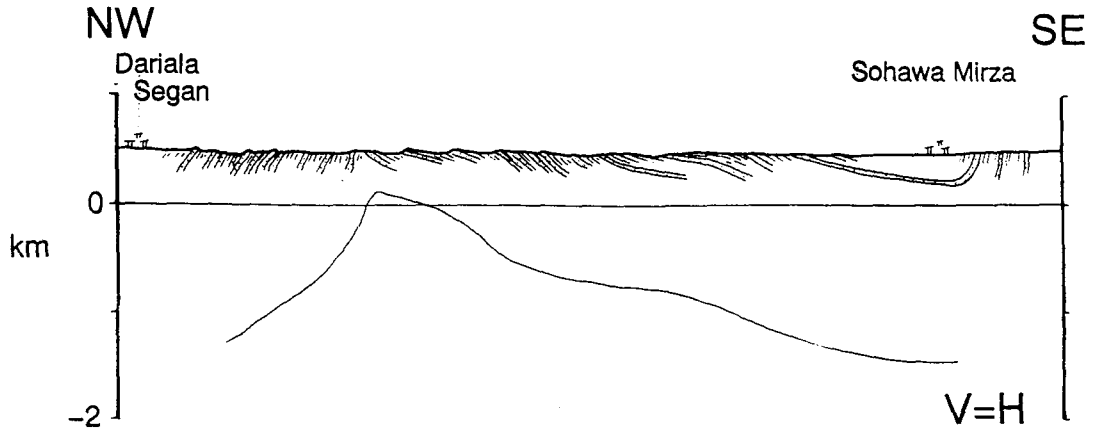


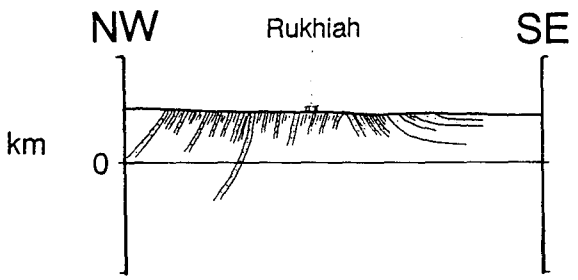
Fig.7.30. INTERPRETATIONS OF SEISMIC TIME SECTIONS 7 & 8, INCORPORATING SURFACE DATA.

(Ornamentation as before)

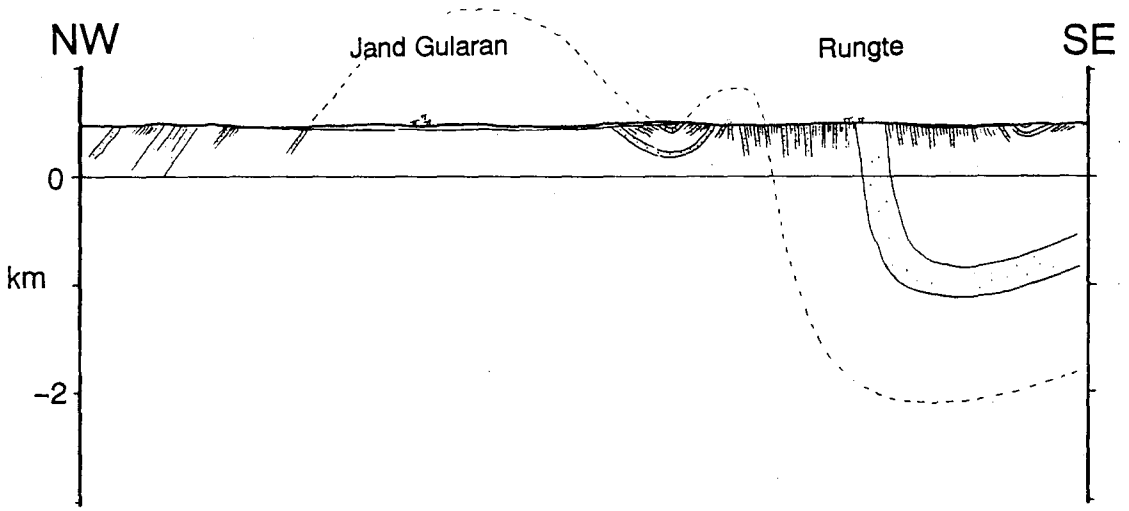
1) Near Jhelum River



2) At Rukhiah



3) At Jand Gularan, just E of the GT road



4) Near GT road

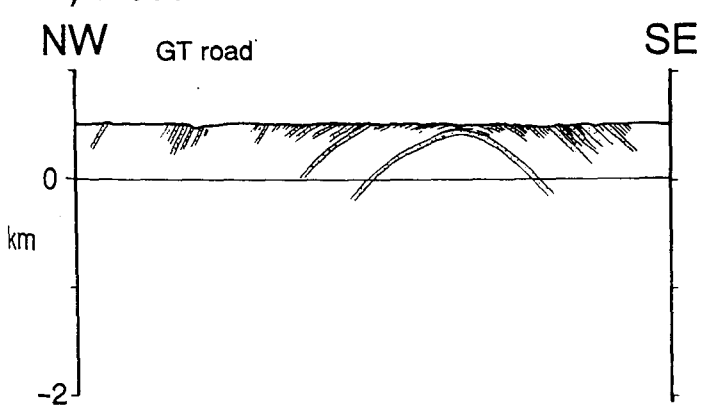


Fig.7.31. ALONG STRIKE VARIATIONS IN SURFACE STRUCTURE AT QAZIAN.

Sections 1–4 illustrate the changing fold profile from NE to SW.

3). The characteristic pop-up geometry only forms towards the east (Fig.7.29, line 5 and Fig.7.30, line 7). This variation is controlled by the increasing uplift on the Domeli Thrust towards the west.

Exploratory wells passed through 2060m of lower Siwalik Group and Rawalpindi Group molasse before reaching the 400m thick Cambrian-Eocene series and finally the Salt Range Formation at about 2460m. Basement reflectors arrive at 3.0secs TWT which indicates the presence of at least 2500m of salt.

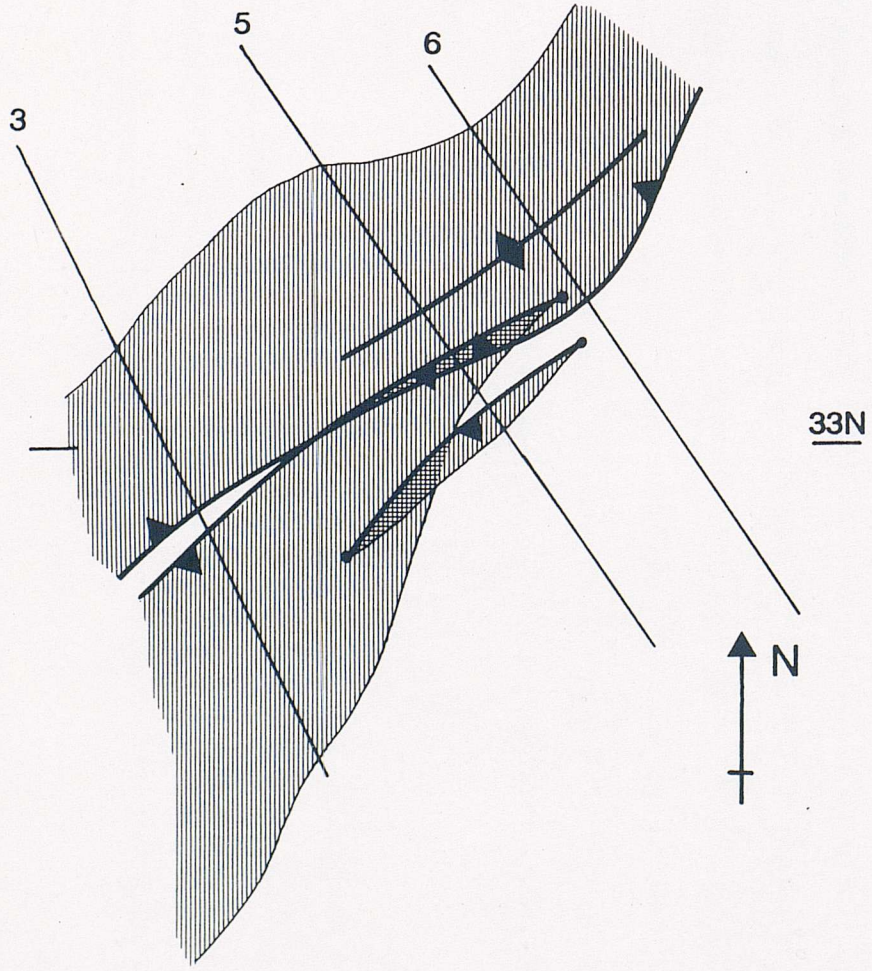
7.3.5. Domeli.

The Domeli structure is the most prominent feature of the Potwar Plateau, forming a ridge that rises 300m above the plateau in places, to summits 800m above sea-level. To the east of Domeli village, Eocene nodular limestones are exposed at the surface having been uplifted approximately 5km above their regional level. The amount of uplift gradually lessens as one moves northeastwards along the axis of the structure, and thus one is able to observe the changes in structural style with depth by walking down the plunge of the structure. For most of its length, the Domeli ridge is marked by a thrust that dips at 20-25°NW. This is inferred from the dip of the Siwalik Group molasse to the northwest of the ridge (Fig.7.23) and supported by seismic lines that traverse the ridge (Fig.7.28, line 3; Fig.7.29, lines 5 & 6 and Fig.7.30, line 7).

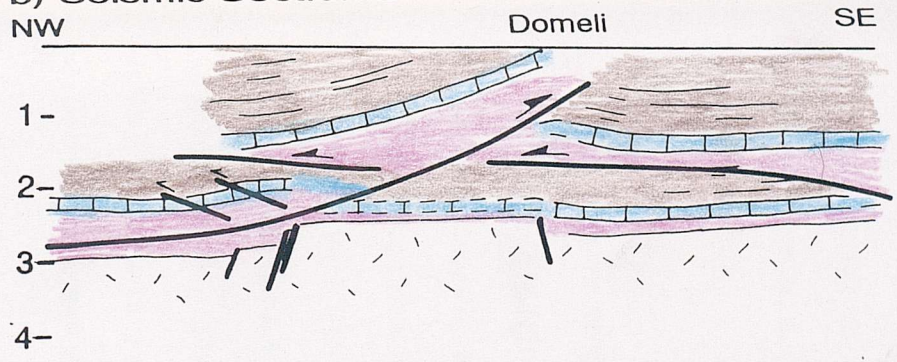
The hangingwall geometries exposed on the ridge in the Eocene limestones and the Rawalpindi Group molasse are quite complex however, and do not display either a simple hang-ingwall flat or ramp geometry. Instead a combination of fore- and backthrusts are linked along strike by transfer zones to describe the crest of a pop-up. At the thrust tips the boxfold geometry predicted at depth in the Ghungrila-Kallar and Qazian structures is observed.

The SE-vergent Domeli thrust carries subvertical, pinkish red/blue-grey-coloured Chinji Formation sediments over dull pink/brown-coloured Dhok Pathan Formation shales and sands at Ruppar. The thrust accommodates 7km shortening (8.8km according to Pennock, 1987) which

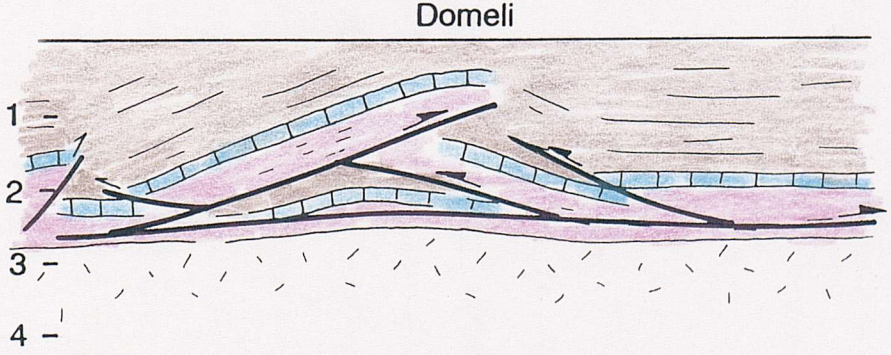
a) Fault Cut–Off Map



b) Seismic Section 3



c) Seismic Section 5



d) Seismic Section 6

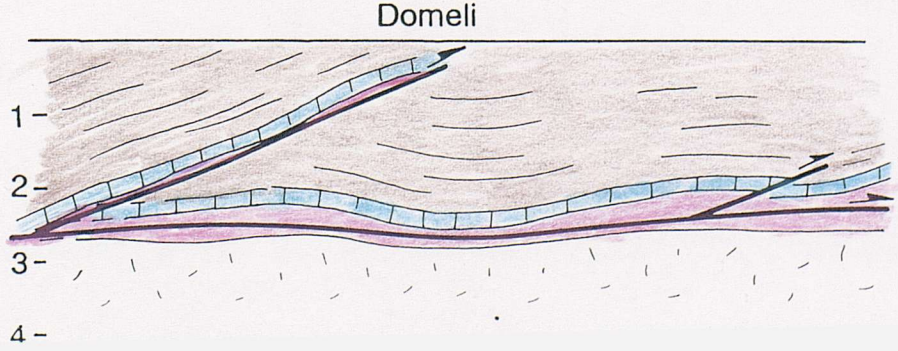


Fig.7.32.

VARIATION OF THE DOMELI STRUCTURE ALONG STRIKE

represents a large part of the total displacement observed within the Potwar Plateau. Towards the northeast, however, this displacement decreases since the structure terminates as a tight fold that plunges at 21°NE.

Towards the southwest the style of thrusting changes. Cut-off maps (Fig.7.27) and TWT contour maps (Fig.7.26) of the Domeli structure, derived from seismic data, show how a back-thrust that initiates between seismic lines 5 & 6 (Fig.7.29) increases in displacement towards the southwest. The effect of this backthrust is to drive a wedge under the Domeli forethrust which uplifts the Domeli Thrust (Fig.7.32b). This uplift has already been mentioned with respect to its effect upon the Qazian structure. Here one finds the evidence for why the Domeli thrust sheet is uplifted towards the southwest. Towards the Salt Range the Domeli Thrust is replaced by the Dil Jabba Thrust which is a NW-vergent structure (Gee, 1981) and probably represents the south-westerly extension of the backthrust beneath Domeli.

7.3.6. Jogi Tilla Thrust/Mahesian.

The Salt Range Thrust, which outcrops along the southern flank of the ENE-WSW trending Salt Range, is the frontal emergent thrust of the Pakistani Himalayas. It is limited to the west by the Kalabagh lineament which has been variously described as a dextral strike-slip fault (McDougal, 1985) and as a lateral ramp (Butler et al, 1987). To the east, displacement on the SRT gradually diminishes and is distributed between the structures of the eastern Potwar, principally the Domeli, Jogi Tilla-Mahesian, Rohtas and Pabbi structures. Topographic analysis alone would suggest that the SRT extends eastwards around the Chambal ridge as the Jogi Tilla and Rohtas ridges (Fig.7.23). However, the continuity of geological structures at depth is more complex.

Estimated displacement on the SRT decreases towards the east. Leathers (1987) calculated that displacement on the SRT was 37km in the western Potwar Plateau-Salt Range, but Baker (1987) could account for only 19-23km in the central Potwar Plateau-Salt Range. In the eastern Potwar Plateau, where shortening is not restricted to the SRT but is accommodated in

a more diffuse manner, Pennock (1988) proposes that 23.1km shortening has occurred. This is in close agreement with this present work. The similarity between the shortening estimates from the central and eastern Potwar Plateau–Salt Range sections implies that the change in structural style observed between the two sections was not in response to lateral pinning, but to other factors such as lateral stratgraphic variation (discussed in chapter 8).

The majority of the displacement on the SRT in the central Potwar (19–23km according to Baker (1987)) is accommodated by three structures in the eastern Potwar – Domeli, Jogi Tilla/Mahesian and Rohtas – all of which splay from the same Eocambrian detachment horizon as the SRT. Pennock (1988) provides the following shortening estimates : Domeli (8.8km), Mahesian (4.2km) and Rohtas (4.5km).

Detailed surface and subsurface mapping shows that the along strike continuation of the Jogi Tilla thrust is the Mahesian structure (Fig.7.33). Displacement on the SSE–vergent Jogi Tilla thrust decreases rapidly as it ramps laterally towards the east. This ramp is clearly seen on seismic strike lines (Fig.7.34, line S). The trend of the Jogi Tila thrust swings from NE–SW to N–S at this ramp (Figs.7.23 & 7.33). At this point the structure has the form of a symmetrical anticline that has been breached by both foreland– and hinterland–vergent, small displacement thrusts (Fig.7.30, line 7). Further east at Mahesian, however, the NE–SW trend is resumed as a single backthrust becomes dominant (Fig.7.30, line 8). In the extreme northeast, the surface expression of the structure is the strongly NW–vergent Lehri fold (Fig.7.35, line 9). This is interpreted as a tipfold to a NW–vergent thrust that splays off the main Mahesian thrust at depth. The S–shaped surface trace of the Jogi Tilla/Mahesian structure is, therefore, the result of a change in thrust polarity at depth. This polarity flip is accommodated by a complex zone of transfer faults that are observed both at the surface and on seismic profiles, and is best depicted on a fault cut–off map (Fig.7.33).

Jogi Tilla rises 700m above the Punjab plain and affords exposure of the full cover se–quence. The Cambrian–Palaeocene part of this sequence is very thin (approx. 450m) at this point

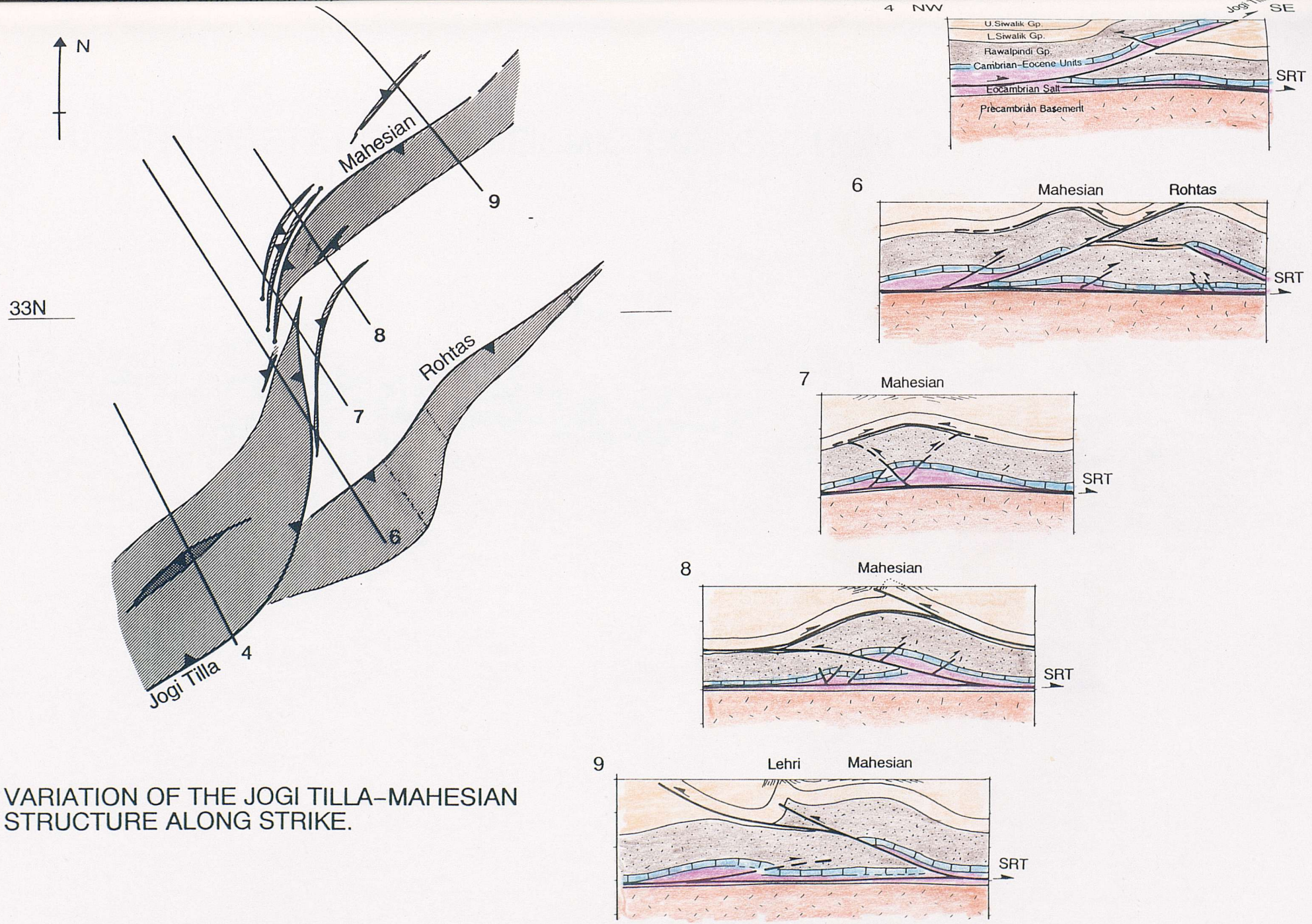
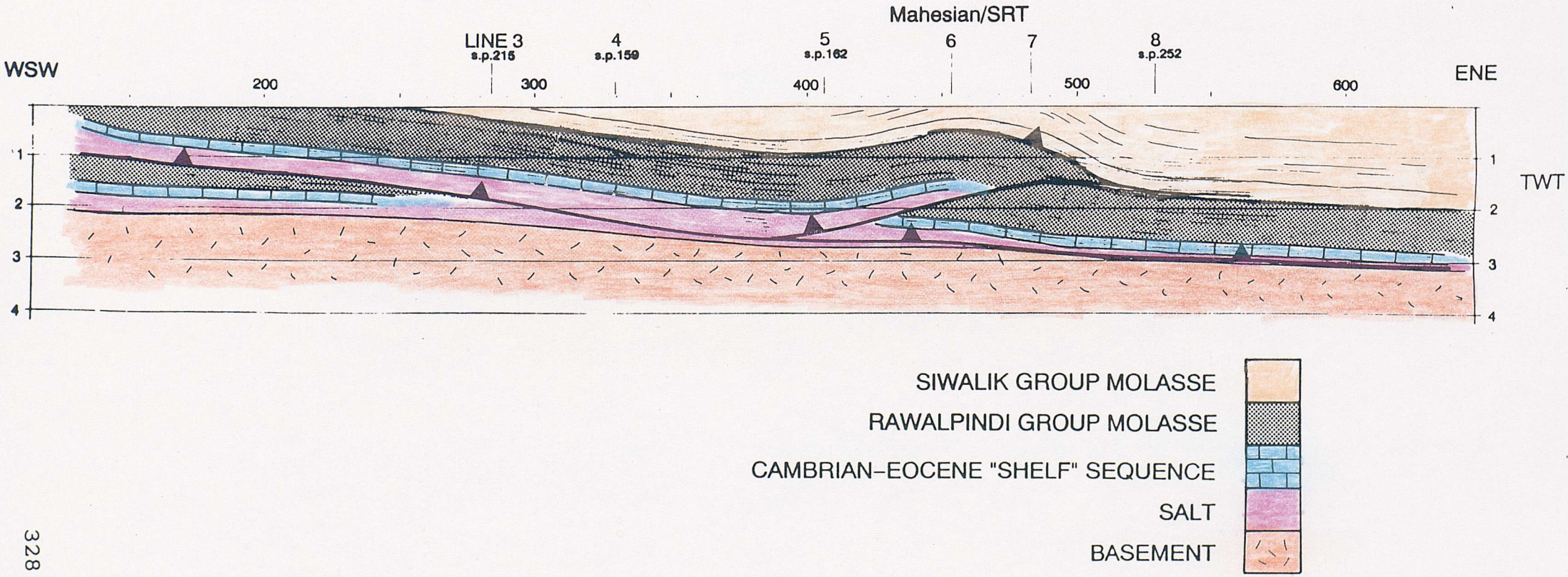


Fig.7.33. VARIATION OF THE JOGI TILLA–MAHESIAN STRUCTURE ALONG STRIKE.

Fig.7.34. INTERPRETATION OF SEISMIC TIME SECTION "S"

STRIKE LINE



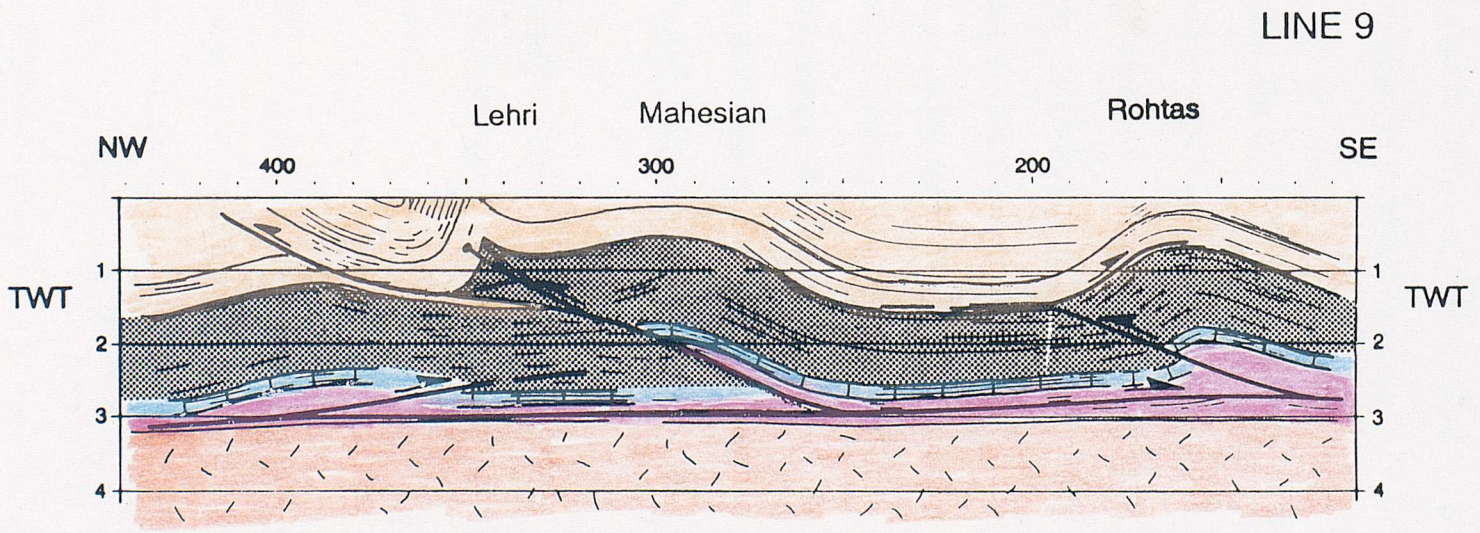


Fig.7.35. INTERPRETATION OF SEISMIC TIME SECTIONS 9, INCORPORATING SURFACE DATA.

(Ornamentation as before)

due to the action of base Permian, base Palaeocene and base molasse unconformities. The presence of salt exposures at the base of this sequence confirms that the Jogi Tilla thrust has ramped from a detachment within the Salt Range Formation. The hangingwall sequence is juxtaposed against Dhok Pathan Formation molasse in the footwall. Towards the northeast the stratigraphic separation across the fault decreases rapidly.

On the southeastern flank of Jogi Tilla the entire Cambrian sequence is exposed. Locally, this sequence dips southeastward in the hangingwall ramp of the Jogi Tilla Thrust or "SRT". At the crest of the Jogi Tilla ridge Cambrian salt pseudomorph beds of the Baghanwala Formation form a veneer above the resistant bluff of Jutana Formation Magnesian sandstones, which dip at 30–35°NW. Descending the northwestern flank of the ridge bedding dips increase to approximately 60°NW before gradually flattening into the Kotal Kund syncline (Fig.7.23). The Cambrian series is unconformably overlain by the Palaeocene Patala Formation (Gee, 1981). This unit is coal–bearing and is still being mined from adits. The Eocene Margalla Hill limestone, which lies with conformity upon the Patala Formation, is unconformably overlain by the Miocene Rawalpindi Group molasse series. The unconformity is marked by a thin conglomerate.

At the northern limit of Jogi Tilla, however, late uplift on the Rohtas structure has caused backsteepening of the beds in the hangingwall to the Jogi Tilla thrust. In places bedding is overturned and dips at 70°SE.

The Mahesian fold exposes Chinji Formation shales and sands in its core. Four traverses of the northwest–vergent anticline were mapped (Figs.7.23 & 7.33), revealing a consistent fold profile along strike. The axis is tight. Bedding dips at 65–75°NW on the steeper northwestern limb and at 15–30°SE on the southeastern limb. The fold becomes more open and gentle as it plunges towards the northeast. In this region shortening is accommodated by the northwest–vergent Lehri fold. Both the Mahesian and Lehri folds are fault–propagation tipfolds.

7.3.7. Rohtas.

Seismic profiles across the Rohtas structure show reflectors rising towards the northwest indicating the presence of a southeast–dipping backthrust at depth (Fig.7.29, line 6; Fig.7.30, lines 7 & 8 and Fig.7.35, line 9). This is not obvious from the surface geology (Fig.7.23). A NW–SE traverse from Rohtas fort along the Khan river reveals an angular boxfold (near Fig.7.30, line 8). In its axial zone bedding dips at 15–20°NW for 1–1.5km and this zone is flanked by domains of steeply–dipping bedding. In the north, bedding dips at 60–65°NW and in the south bedding dips at 80–90°SE.

In a second parallel traverse, 8km along strike to the southwest (near Fig.7.30, line 7), the boxfold geometry is gone and the fold has been breached by a southeast–vergent thrust which over–rides overturned beds in its footwall. The Lower Siwalik hangingwall sediments dip at 35–45°NW.

The northeastern tip of the Rohtas fold is observed at the G.T. road section (Fig.7.23; near Fig.7.35, line 9) where the fold is open and symmetrical at the surface. Bedding dips gently at 18–22° on both fold limbs. At this locality the anticline plunges at approximately 10–15°NE.

The disharmonic deformation of the Siwalik sediments observed at the surface and the Cambrian–Eocene shelf carapace at depth is characteristic of the eastern Potwar region. The two data sets confirm the existence of the intramolasse detachment across the entire eastern Potwar from the Riwat Thrust to the Rohtas anticline. The precise stratigraphic horizon at which the detachment occurs is not known. From exploration drilling results it is known that numerous overpressured horizons exist within the Siwalik Group sediments. The overpressures result from the rapid sedimentation rates in the Himalayan foreland basin and the consequent inadequate dewatering of the sedimentary pile. Thin–sections of the Siwalik sandstones show clear evidence of high fluid pressures (M. de Freitas, *pers. comm.*). Seismic evidence suggests that the detachment lies at a regional level of approximately 2 secs. TWT which is equivalent to approximately 3km depth, although the detachment is uplifted at culminations. Construction of cross–sections based on field data constrains the detachment to lie within the basal portion of the Chinji Forma–

tion, which is compatible with the figure obtained from depth conversion of the seismic data.

The development of the Rohtas deep backthrust and shallow forethrust tipfold postdates both the Jogi Tilla thrust and the Mangla fold. The presence of the Jogi Tilla thrust has constrained the lateral propagation of the Rohtas thrust. As mentioned earlier (section 7.3.6) uplift on the Rohtas structure is responsible for overturning of beds at the northeastern tip of the Jogi Tilla ridge and the extra topography of this ridge. The existence of the WNW–ESE trending Mangla fold to the northeast of Rohtas constrained the latter's northeastward propagation. The formation of the Mangla fold has been dated palaeomagnetically by Johnson et al (1979) at 2.7–1.5Ma whilst the Rohtas fold did not grow until 1.2–0.4Ma (*op. cit.*)

7.3.8. Pabbi.

The formation of the Pabbi fold resulted from uplift above a gently NW–dipping forethrust during the period 1.2–0.4Ma (Johnson et al, 1979). It is therefore the youngest structure in the Pakistani Himalayas. To the south of Pabbi lies the undeformed Himalayan foreland.

The stratigraphy at this locality, the most southeasterly corner of the Potwar Plateau, is dominated by Upper Tertiary molasse sediments. The Lower Tertiary strata are absent in this area due to a period of erosion represented by the basal molasse unconformity. Thus the Cambrian Jhelum Group sediments and Miocene Rawalpindi Group molasse are juxtaposed (unpublished commercial data).

The Pabbi Hills Thrust ramps from the basement–cover detachment within the Eocambrian salt, through the thin Cambrian sequence and terminates within the molasse. Uplift on the Pabbi structure is approximately 500m and therefore relatively minor compared with the other structures of the eastern Potwar. Only Upper Siwalik sediments are exposed in the Pabbi Hills.

The anticline verges southeast, with Dhok Pathan Formation strata dipping at 40°SE on the steeper southeastern limb and at 12–16°NW on the northwestern limb. The hinge zone is broad

and open.

7.3.9. Summary of the SE. Potwar.

1. Deformation of the SE Potwar occurred between 3.4Ma and the Present (Johnson, Raynolds & Burbank, 1986).

2. Shortening estimates for the central (19–23km, Baker (1987)) and eastern (23.1km, Pennock (1988)) Potwar are similar. The structures of the eastern Potwar accommodate in a diffuse manner the shortening accommodated by the SRT in the central Potwar.

3. The structures of the eastern Potwar nucleate above a salt detachment (the SRT or Potwar–Salt Range Detachment). Halokinesis during Himalayan compression initiated salt pillows that grew during molasse sedimentation. Maximum sediment thickness occurs in the intervening broad, open synclines.

4. Continued shortening was accommodated by pop–up formation. The pop–up bounding thrusts either terminate as box–shaped tipfolds or flatten onto an intramolasse detachment. The level of the detachment is determined by formation pressures. Rapid burial in the foreland basin meant that dewatering was incomplete and high fluid pressures resulted.

5. Further shortening was then accommodated by either :

a) through–going low angle thrusting which breached the intramolasse detachment and carried the Cambrian–Eocene sequence to the surface, e.g. Domeli, or

b) the merging of pop–up culminations. As two pop–ups propagate towards each other along the intramolasse detachment the intervening broad, flat–bottomed syncline is gradually tightened and uplifted until eventually it is completely eroded (Fig.7.29, line 6).

CHAPTER 8.

AN EVOLUTIONARY MODEL FOR THE WESTERN LIMB OF THE HAZARA-KASHMIR SYNTAXIS.

In the previous chapter the available field and seismic data from southeastern Hazara and the eastern Potwar was described. Here a model for the evolution of this area, which comprises the western limb of the Hazara–Kashmir Syntaxis, will be presented. The model charts the propagation of the Himalayan thrust wedge in northern Pakistan from NNW to SSE during the Neogene.

Firstly, the factors leading to the growth of the NW Syntaxis, which comprises numerous smaller syntaxes e.g. the Hazara–Kashmir, Besham and Nanga Parbat Syntaxes, will be discussed. Secondly, the theoretical mechanisms of HKS formation will be described. Finally, the evidence from the western limb of the HKS will be used to constrain the deformation phases and, hence, the evolutionary model for the region.

8.1. FACTORS CONTROLLING THE GEOMETRY OF THE HIMALAYAN OROGENY.

The zone affected by Himalayan orogeny extends over much of southern central Asia. The Himalayan thrust wedge is characterised by three segments – the N–S trending Kirthar and Sulaiman Ranges (Pakistan), the Main Himalayan arc and the N–S trending Arakan Yoma Range (Myanmar). These segments link along strike at two major syntaxes, the NW Syntaxis and the NE Syntaxis (Fig.8.1).

This geometry is the product of three principal factors :

1) The shape of Greater India and the convergence vector dictated that initial India–Asia collision occurred at the proto–NW Syntaxis. The age of collision becomes progressively

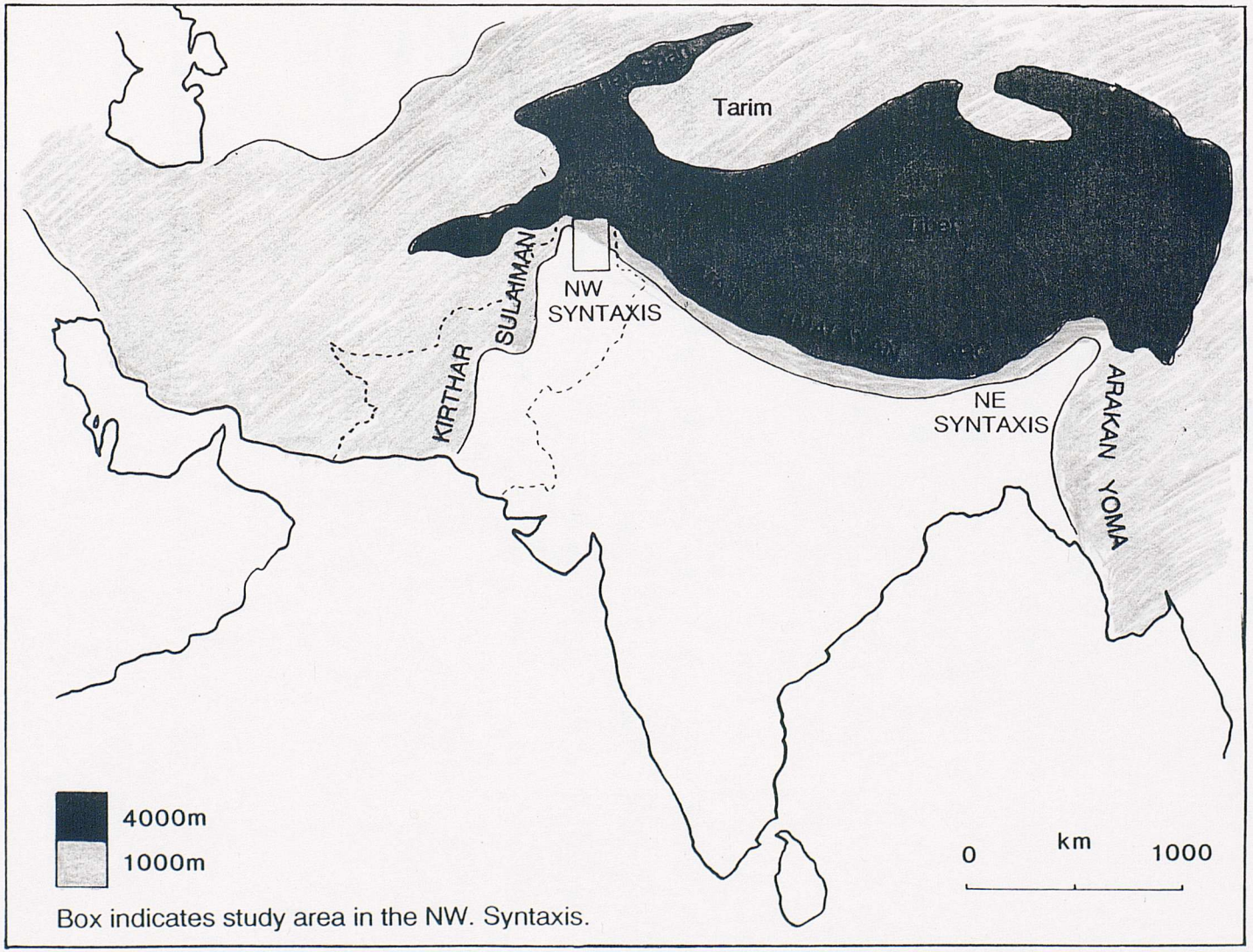


Fig.8.1 MAP SHOWING THE EFFECTS OF HIMALAYAN OROGENY IN SOUTHERN ASIA.

younger towards the east along the ITSZ. The early development of a thrust wedge in the NW Syntaxis will have laterally inhibited the propagation of the subsequent Main Himalayan arc. Thrusts propagating laterally westward will have encountered the NW Syntaxis and been pinned, a situation that once initiated has persisted to the present day (Treloar et al, in press; Treloar & Coward, 1991).

As well as dictating the position of initial collision the shape of India has controlled the nature of collision on India's respective passive margins. The northern Neotethyan margin has been the site of oblique/frontal overthrusting since collision, with convergence rates varying from *ca.*50mm/yr in the west to *ca.*70mm/yr in the east as one moves away from the pole of rotation between India and Asia. The northwestern margin of India is almost perpendicular to the Neotethyan margin and has suffered sinistral transpression since collision.

2) The geometry of oceanic subduction prior to collision. It is well-documented that Neotethys was subducted beneath the southern margin of Asia during the Mesozoic leading to the generation of the Gangdese or Transhimalayan plutonic arc (Argand, 1924; Gansser, 1964; Coulon et al, 1986). The northward drift of India closed the Neotethys Ocean and juxtaposed the Andean-type margin of Asia against the mature passive margin of India. The subduction polarity dictated that the Indian margin was subducted beneath Tibet and that a S-vergent shear couple was established. This fact led to the formation of the southward convex Main Himalayan arc.

3) The relative strength of India compared to Asia at collision. India is a major fragment of Archean continental shield. Southern Asia comprises a mosaic of continental blocks that were rifted from Gondwana and accreted to Asia during the Mesozoic "Cimmeride" orogeny (Sengor, 1990). Upon India-Asia collision, India continued its northward drift and after initial thickening of its passive margin began to indent into the Cimmeride mosaic. Dewey, Cande & Pitman (1989) suggest that in the western Himalayas 50% of Himalayan shortening was accommodated by thickening of Tibet and 50% by imbrication of the Indian crust. This means

that major underplating did not occur but rather India has indented into Asia – a suggestion supported by palaeomagnetic data from Tibet indicating northward displacement of the ITSZ by 1980km (Dewey et al, 1988). Indentation has accentuated the NW and NE Syntaxes by generating the two N–S trending segments of the Himalayan system.

The development of the NW Syntaxis therefore is a consequence of the fact that India is an irregularly–shaped, under–riding indentor. Throughout Himalayan orogeny the NW Syntaxis was a site of thrust interference and small–scale syntaxial development. The Hazara–Kashmir Syntaxis is one such structure.

8.2. THE FORMATION OF THE HAZARA–KASHMIR SYNTAXIS.

The HKS is a region that has been uplifted by over 2km during the Quaternary period, implying an average uplift rate of 2mm/yr. Erosion of the core of the uplifted dome has revealed the imbricated nature of the HKS. The overturned Hazara Detachment can be traced along the western limb of the HKS juxtaposing the Murree Formation molasse sediments in its footwall with progressively older Eocene to late Precambrian sediments in its hangingwall. In the apex of the HKS the Hazara Detachment passes into an extensive NNE–NE trending mylonite zone. In its footwall it links with the Indian MBT which swings southeastward. The Muzaffarabad and Kotli Thrusts testify to imbrication in the footwall of the MBT.

Several mechanisms have been proposed to explain the formation of the HKS. They fall into three main categories.

1) Firstly, syntaxes or re–entrants may develop in unidirectional thrust systems where the tectonic transport direction is relatively constant through time (Wadia, 1928). This category comprises two main groups :

a) **The formation of large scale antiformal stacks.** One can generate extra uplift in a certain area by either a concentration of shortening in the form of an antiformal stack (Butler, 1982) or a thickening of the allochthonous section along strike by thrust ramping or by

stratigraphic thickness variations. Erosion of the uplifted dome will reveal an arcuate pattern.

b) **The interaction of basement highs with the advancing thrust front.** In contrast to the above mechanisms which are genetically arcuate, one can also envisage a unidirectional thrust system in which an initially linear thrust front is deformed into an oroclinal geometry by the the indentation of a rigid mass, e.g. a basement high in the advancing foreland. This situation might be expected in a passive margin setting where the basement–cover contact is offset by faults. Wadia (1932) recognised this geometry in Pakistan and proposed indentation as a model for the formation of the HKS, a model which could be extended to the Mianwali and Sibi re–entrants also. Clearly, large rotations about a vertical axis are predicted by Wadia's model that are testable by palaeomagnetic methods. In addition, one would predict basement at shallow depths under the HKS according to this model. In fact we observe basement descending at *c.*3 degrees northeastwards under the HKS such that basement lies at greater depth under the HKS than under the Potwar Plateau. Indentation therefore can be ruled out as a mechanism of HKS formation. The development of antiformal stacks is common within thrust belts. The scale and aspect ratio of the HKS with respect to the transport directions observed in Hazara and the orientation of those thrusts that are exposed in the core of the HKS make such unidirectional stacking an unlikely mechanism of HKS formation.

2) Secondly, they may develop due to a change of tectonic transport direction or rotation through time. This category comprises models in which a single transport direction is operative at any one time but this direction varies through time. This model has been forwarded for the formation of the HKS (Bossart et al, 1988) and the Ponga Arc in N.Spain (Alvarez–Marron & Perez–Estaun, 1988). They argue that the syntaxis is the result of an anticlockwise swing in thrust transport direction from SW– to SE–vergent which causes the originally NW–SE trending thrusts to be folded into the present–day orocline.

This model has been modified (Bossart, Ottiger & Heller, 1989; 1990) to account for palaeomagnetic results that show that the Hazara thrust wedge has not endured the predicted anticlockwise rotation but that the Kashmir thrust wedge has been rotated in a clockwise

direction by 45° (Fig.8.2). Their new model proposes that the early transport directions on both sides of the HKS reflected the plate convergence vector (NNW–SSE) but that the thrust wedge on the eastern side of the syntaxis has been passively rotated by 45 degrees in a clockwise direction such that the transport direction appears to be N220E. The suggested reason for this rotation is pinning of the Main Himalayan thrust wedge at the Indus–Kohistan Seismic Zone (ITSZ) to the northwest of the HKS (Treloar et al, in press).

3) Thirdly, they may form in bidirectional thrust systems due to the interference of convergent tectonic transport directions that are active simultaneously. This category recognises two independent transport directions throughout the evolution of the NW Syntaxis and that local transport directions vary considerably from the plate convergence vector. Treloar et al (*op.cit.*) propose the interference of two concurrent, but convergent, thrust wedges. The SSE–vergent Hazara wedge and the SW–vergent Kashmir wedge interfere to produce extra uplift in the HKS. Rotations within the Kashmir wedge are attributed to pinning of SW–vergent thrusts as they attempt to propagate under the Hazara wedge. Treloar et al (*op.cit.*) note that the degree of rotation will lessen as one descends the thrust pile. This is in opposition to the data of Bossart, Ottiger & Heller (1990) who observed a 45 degree clockwise rotation of the strata in the hangingwall of the Kotli thrust which is dated at 2.8–1.5Ma.

The convergence in transport direction is inherent in the understanding of how the NW Syntaxis as a whole has evolved (section 8.1.). The original shape of India at collision, which dictates the trend of the mountain front, established the NW Syntaxis as the zone in which the Main Himalayan and Pakistani thrust wedges interfere. Each wedge is inhibited and pinned by the other producing passive rotations on a large scale, in an anticlockwise direction on the western side and clockwise on the eastern side of the HKS. These passive rotations may be locally disguised by small scale active rotations related to the displacement of individual thrust sheets, e.g. the eastern Salt Range (C.Jackson, *pers.comm.*).

Field data from the western limb of the HKS supports the simultaneous

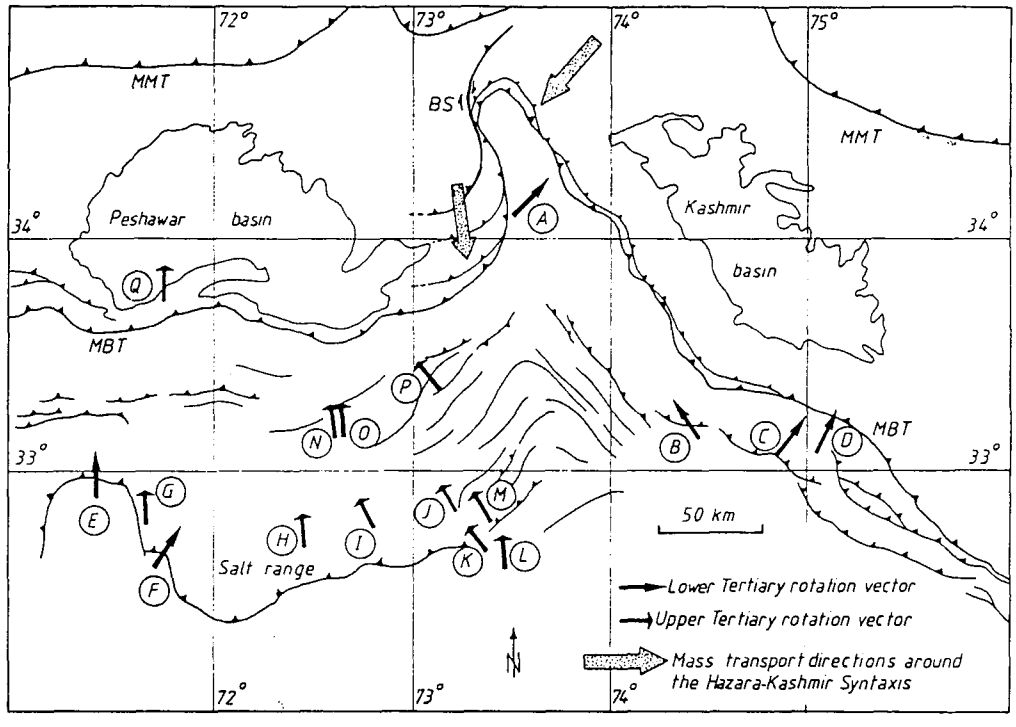


Fig.8.2 DISTRIBUTION OF PALAEOMAGNETIC VECTOR AZIMUTHS AROUND THE HAZARA–KASHMIR SYNTAXIS, from Bossart, Ottiger & Heller (1990).

Note that the vectors have been rotated anticlockwise on the western limb of the HKS and clockwise on the eastern limb of the HKS.

operation of two transport directions. During Pliocene times SW–vergent compression in the footwall of the MBT in Kashmir produced crossfolds in the hangingwall of the Hazara Detachment whilst SSE–vergent deformation was occurring in the S. Potwar. It is suggested therefore that the SSE–vergent Hazara thrust wedge and the SW–vergent Kashmiri thrust wedge can be considered as independent thrust systems until the Pliocene, at which time they began to interfere and uplift of the HKS commenced. Prior to this time the zone of active interference lay to the NNE of the HKS.

8.3. DEFORMATION HISTORY OF THE WESTERN LIMB OF THE HKS.

The Hazara–Potwar thrust wedge which lies on the western limb of the HKS has propagated south–southeastwards (N160E) during the last 30Ma. This period includes the post–metamorphic imbrication of internal zones (Williams, 1990; Coward et al, 1988) as well as the brittle deformation of the external zones in the Hill Ranges and the Potwar Plateau (Yeats & Lawrence, 1984; Johnson et al, 1986; Yeats & Hussein, 1987; Lillie et al, 1987). Deformation of the area between the Nathia Gali Thrust and the Salt Range Thrust (described in chapter 7) occurred towards the end of this period. The duration of active deformation between the NGT and the SRT is poorly constrained. Izatt (1990) suggests that the MBT became active at *ca.*8.5Ma and this was preceded by deformation of its hangingwall. Dewey, Cande & Pitman (1989) provide estimates of the total orogenic shortening accommodated along the India/Eurasia convergence zone. In the Western Syntaxis they partition this shortening between the Tien Shan (390km), Tibet (737km) and the Himalayas (1188km). From this data, a time–averaged estimate of the Himalayan shortening rate during the last 45Ma is 26mm/yr. The estimated shortening in the footwall to the NGT is 240km and thus, the deformation of this region began at *c.*9Ma. This fits well with Izatt's (1990) estimates.

By 9Ma the Potwar basin had already subsided to accommodate 2–3km of Rawalpindi Group molasse sediments – subsidence increasing towards the northeast and north. Since the Upper Miocene a further 3–4km of Siwalik Group sediments have been accommodated

implying subsidence rates of 0.4–0.5mm/yr. The axis of the foreland basin today trends parallel to both the Main Himalayan arc and the Pakistani thrust wedge. Basement dips, therefore, towards the northeast in Jammu but towards the north–northwest in the western Potwar and Kohat. In the NE.Potwar, the basin is influenced by both thrust loads and dips at *ca.*3.5° towards N030°E. This basin geometry probably existed throughout the 9–0Ma period during which the footwall to the NGT has deformed, althouigh the Potwar Basin may have trended NW–SE prior to the rapid advance of the SSE–vergent Hazara thrust wedge.

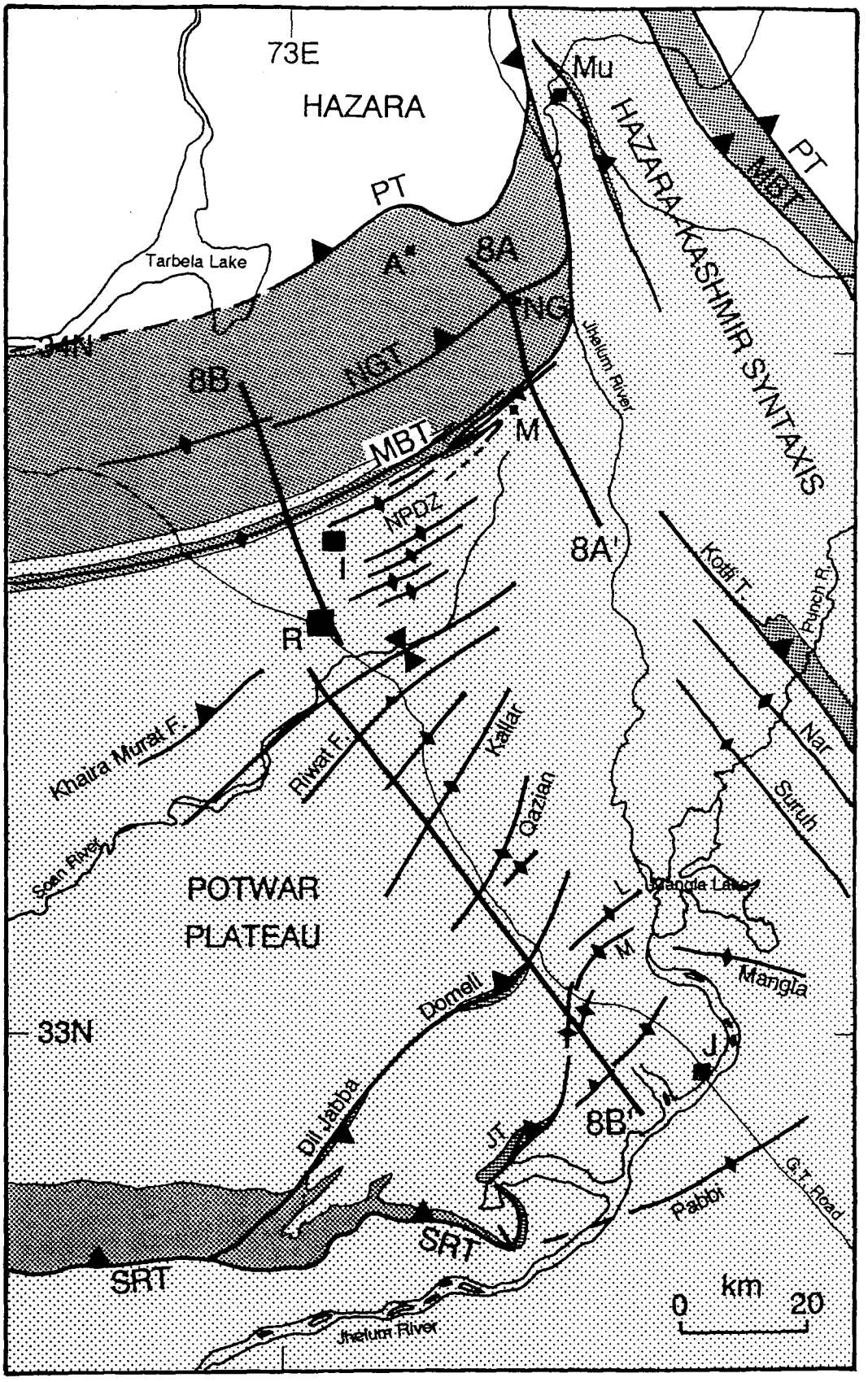
The deformation of the external zones occurred therefore within a rapidly subsiding molasse basin which deepened towards the northeast as well as parallel to the transport direction. The implication of this is that in order to remain at the same topographic level a SSE–vergent thrust must ramp laterally towards the east. This is thought to occur in Hazara and has been proposed by Butler et al (1987) in the SE.Potwar.

During the deformation of the Hill Ranges and the Potwar Plateau two major detachments have been utilised – the Hazara Detachment (or MBT) and Potwar–Salt Range Detachment (or SRT). These detachments branch where the MBT ramps from the basement–cover contact to the north of the HKS (Figs.8.3 & 8.4). Correlation of these structures around the HKS is problematic. Commonly the MBT in India is equated with the Murree Thrust/Hazara Detachment in Hazara (Burbank, 1983; Johnson et al, 1986; Lillie et al, 1987; Coward et al, 1988) but these structures have different ages, the Hazara Detachment being active until *ca.*6Ma compared to the MBT in India which is dated at *ca.*3Ma (Burbank et al, 1986). The SRT was active therefore at the same time as the Indian MBT.

The evolution of a section along the western limb of the HKS is discussed in the following sections (Figs.8.3 & 8.4). **Its deformation history can be divided into four main phases** :–

1) The **imbrication of the hangingwall to the Hazara Detachment**. During the period

Fig.8.3 SIMPLIFIED STRUCTURE OF THE EASTERN POTWAR AND SOUTHERN HAZARA, showing section lines 8A–A' & 8B–B'.



External Zones — Pre-Himalayan sediments; Himalayan molasse

PT, Panjal Thrust NGT, Nathia Gali Thrust
MBT, Main Boundary Thrust SRT, Salt Range Thrust

A, Abbottabad; I, Islamabad; J, Jhelum; M, Murree;
Mu, Muzaffarabad; NG, Nathia Gali; R, Rawalpindi.

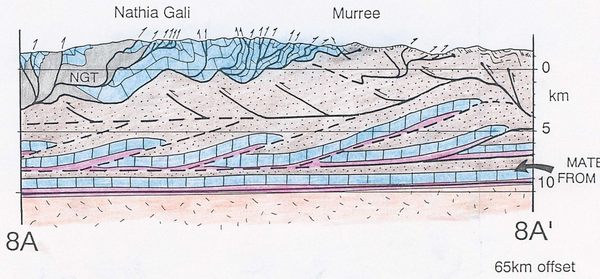
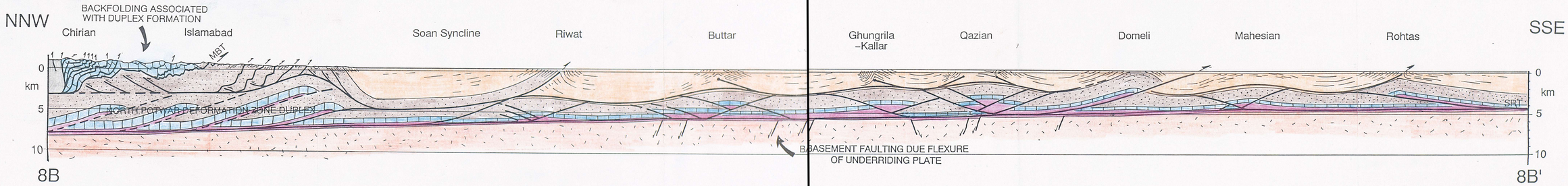


Fig.8.4 STRUCTURAL CROSS-SECTION ACROSS THE HILL RANGES AND THE EASTERN POTWAR.

(incorporating Hill Ranges data from Izatt, 1990)

*c.*36km SHORTENING IN FOOTWALL TO NGT
>80km TRANSLATION ON THE MBT
*c.*80km SHORTENING IN FOOTWALL TO MBT i.e. the NPDZ duplex
*c.*33km SHORTENING IN FOOTWALL TO NPDZ DUPLEX



For location of section see Figure 8.3.

*ca.*9–8Ma the Hill Ranges between the NGT and the Murree Thrust were shortened by approximately 36km.

2) **Translation on the Hazara Detachment.** Between 8–6Ma the MBT or Hazara Detachment hangingwall was displaced south–southeastwards by 85–90km across the molasse basin.

3) **Collapse of the footwall to the Hazara Detachment to produce a duplex.** This duplex has been inferred from seismic data in the northern Potwar where it is known as the NPDZ duplex. Its northward continuation under the Hill Ranges is assumed. Shortening may have commenced during the previous phase (*ca.*6Ma) but had ceased by 4Ma. Estimated shortening is 30–35km, but this figure is very poorly constrained and may be as high as 80km.

4) **Deformation in the hangingwall to the Salt Range Detachment (SRT).** At 4.5Ma the deformation front of the Hazara–Potwar thrust wedge propagated rapidly to the Salt Range (Leathers, 1987). Since then movement on the SRT has continued and out–of–sequence deformation in the eastern Potwar Plateau has thickened the hangingwall to the SRT. This phase is characteristic of thrust wedge propagation above a salt decollement. Shortening is approximately 33km (Baker, 1987; Leathers, 1987; Lillie et al, 1987; Pennock, 1988; Pennock et al, 1989).

Total minimum shortening during the last 9Ma is approximately 200km. Minimum time–averaged shortening rates are therefore *ca.*2.2mm/yr. Actual shortening is episodic and times of quiesence and rapid displacement are expected during this period. The four phases of progressive deformation will be described in the sections 8.4–8.7.

8.3.1. Imbrication of the Hangingwall to the Hazara Detachment.

The deformation of the Galiat occurred at the same time as Nagri Formation

deposition (*ca.*9–8Ma). Approximately 4km of Murree, Kamlial and Chinji Formation molasse sediments had already accumulated by this time and draped the pre-orogenic "shelf" sequence. None of the molasse sediment is preserved in the Galiat imbricate zone except for exposures of the Murree Formation near the Murree. This suggests that either the Galiat imbricate zone has been deeply eroded so that only the root of each imbricate is preserved, or that a roof detachment, perhaps within or just above the Kuldanna Formation, once separated the imbricate zone from the overlying molasse but has since been eroded. Palaeomagnetic inclination data (Bossart, Ottiger & Heller, 1990) suggest the presence of middle Eocene molasse in the footwall of the MBT. According to the model presented here, this would imply that even older molasse may once have overlain the Galiat imbricate zone. Evidence of this has been eroded and redeposited in advancing foreland basin as Siwalik molasse.

The field outcrop pattern demonstrates that at that time the sole thrust (the Hazara Detachment or MBT) rose from a flat within the Hazara Group close to the basement–cover contact to a flat within the Murree Formation molasse, ramping through the Cambrian–Eocene strata (Fig.8.5). The imbrication of the Galiat can be considered in three phases in relation to the staircase trajectory of the Hazara Detachment.

1) In the "Slate" zone, the sole thrust (the Hazara Detachment) lay within the Eocambrian Hazara Group greywackes close to the basement–cover contact. Between Nathia Gali and Bagnotar, a 1.4–1.6km thick section of Hazara Group greywackes is preserved. The full thickness of the Hazara Group is unknown, but 3000m are recorded at Tanol (Mohammed Ali, 1962). The NGT ramps from the Hazara Group to the Eocene Margalla Hill Formation. It has a minimum shortening of 9.5km in SE. Hazara but it terminates as a tipfold 30km along strike to the southwest at Chirlan (Izatt, 1990).

2) South of Nathia Gali no Palaeozoic strata are exposed in the hangingwall of the Hazara Detachment (MBT) and the sole thrust is inferred to have ramped into the Mesozooic strata. Where the sole thrust lies at the base of the Jurassic Samana Suk Formation

large scale imbricates are developed. This is dictated by the well-cemented, resistant Mesozoic carbonate units (the Samana Suk and Kawagarh Formations) which reach maximum thicknesses of 850m and 600m respectively between Dunga Gali and Changla Gali. These observed thicknesses have been incorporated into the cross-sections through the Galiat but it is possible that intraformational thickening has exaggerated these figures. In the footwall to major imbricates of Mesozoic strata such as the Nathia Gali and Changla Gali Thrusts the basal Palaeocene horizon is often exploited as a slip plane producing small stacks of Lower Tertiary strata, e.g. between Moshpuri and Dunga Gali (Figs.8.5, 7.12a&b).

3) To the south of Khaira Gali the sole thrust ramps to the top of the Patala Formation and only Eocene strata are involved in the imbricates. The absence of molasse sediments within these imbricates argues in favour of a roof detachment at the base of the molasse. The only molasse that could potentially belong to the hangingwall of the Hazara Detachment (MBT) outcrops between Darya Gali and Kuldanna and along strike southwestwards to Tret. If this is the case the basal Murree Formation should have originated at least 90km NNW of Murree and be of similar age to the middle Eocene molasse recorded by Bossart, Ottiger & Heller (1990) in the Jhelum valley. Unfortunately, the sediments have not been dated.

The Galiat imbricate zone is therefore a SSE-vergent wedge lying in the hangingwall of the Hazara Detachment that was active between *ca.*9-8Ma. The area mapped lies close to the overturned exposures of the detachment on the western limb of the HKS. The imbricates show no sign of ramping laterally to the east and so one can assume that the Galiat imbricate zone once extended some way across the core of the syntaxis. Shortening between the NGT and Murree Thrust is estimated at 36km. This compares with a minimum of 20km shortening along strike in the Margalla and Kalachitta Hills (Izatt, 1990).

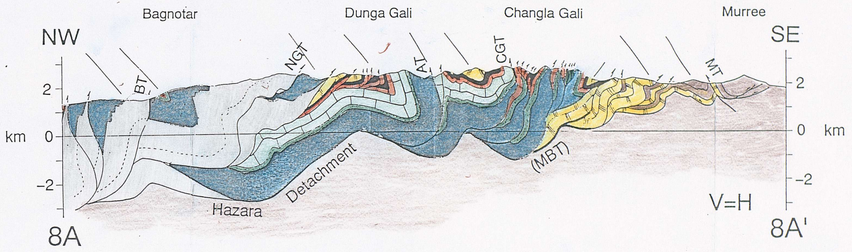
8.3.2. Translation on the Hazara Detachment (MBT).

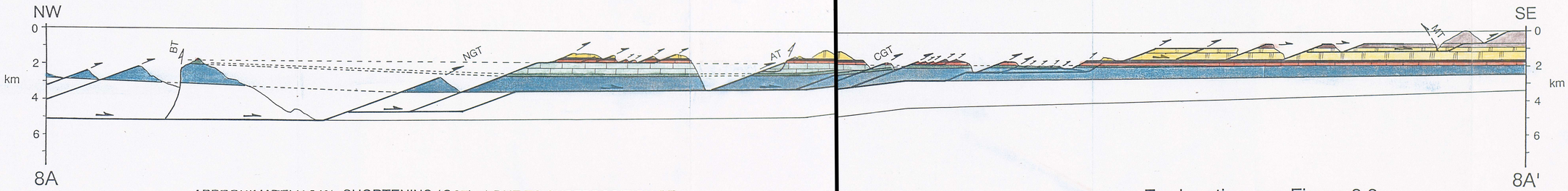
Between 8-6Ma major displacement occurred on the Hazara Detachment (MBT),

Fig.8.5 DEFORMED AND RESTORED CROSS-SECTIONS ACROSS THE GALIAT FROM BAGNOTAR TO MURREE

BT Banotar Thrust
NGT Nathia Gali Thrust
AT Ayubia Thrust
CGT Changla Gali Thrust
MT Murree Thrust

Eocene
Palaeocene
U.Cretaceous
L.Cretaceous
Jurassic





APPROXIMATELY 64% SHORTENING (*C.*37km) DUE TO SUPERPOSITION OF BACKFOLDING UPON EARLY IMBRICATION

For location see Figure 8.3.

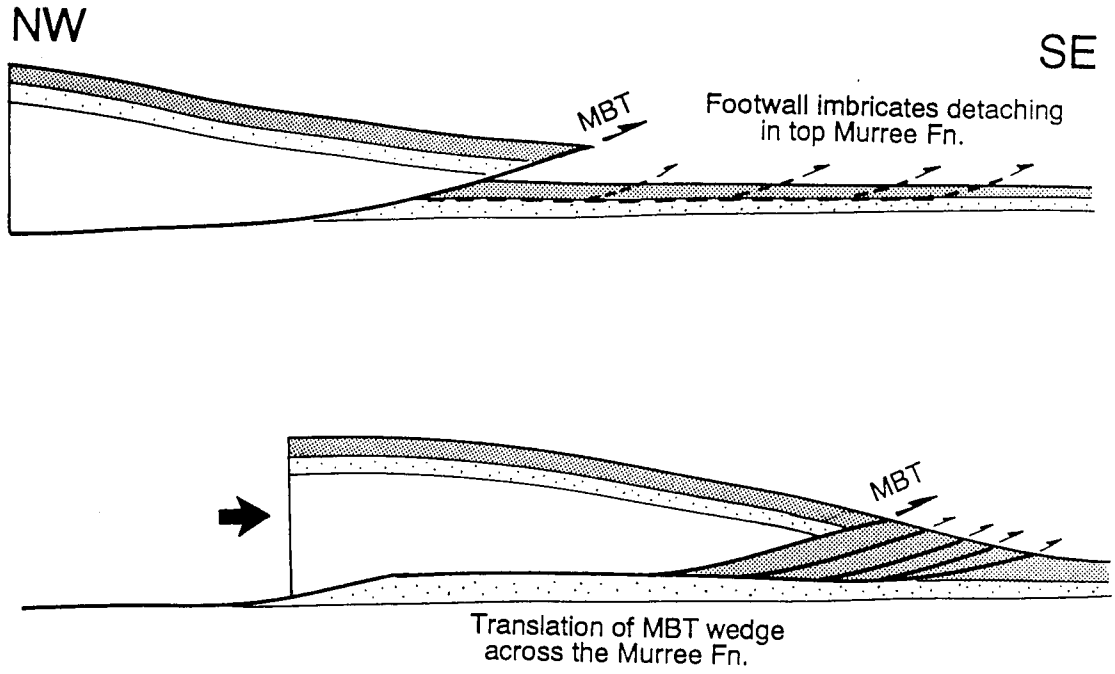
which has been subsequently uplifted and overturned during the growth of the HKS. The detachment is exposed on the western limb of the syntaxis where it carries the Galiat imbricate zone and more internal structures along a footwall flat within the Murree Formation molasse. Displacement was towards the south-southeast (N150°-165°E) (Fig.7.8b). The transition between imbrication of the Galiat zone and the translation of this zone across the molasse basin is associated with the ramping of the MBT from a footwall flat within the Eocene carbonate units to a horizon within the Murree Formation molasse. This juxtaposed the competent and predominantly carbonate wedge against the incompetent and predominantly shaley Murree Formation. Brittle carbonate imbricates were no longer accreted to the wedge tip. Instead, the wedge was driven into the molasse sequence, exploiting the ductile shales of the poorly dewatered molasse series.

L.Siwalik sediments are absent in the footwall of the MBT wedge. The MBT must, therefore, have propagated within the upper Murree Formation, the L.Siwalik strata being either :

a) Uplifted and eroded in its hangingwall (Fig.8.6.(ii)). This process may have occurred by steady-state translation of the wedge across a footwall flat within the Murree Formation by means of a frontal triangle zone (Fig.8.7). This mechanism requires no accretion at the wedge tip. It is proposed that the front of the MBT wedge was marked by a monocline such that molasse strata overlying the MBT detachment horizon were uplifted and eroded in manner resembling the northern limb of the Soan Syncline today. This situation is known in other wedges that have propagated into rapidly subsiding molasse basins, e.g. Sulaiman Ranges (Pakistan), Alberta Foothills (Canada).

b) Imbricated in the footwall of the MBT and subsequently uplifted during NPDZ formation (Fig.8.6.(i)). It is possible that the MBT propagated along a footwall flat within the Murree Formation by footwall collapse of the upper Murree, Chinji and Nagri Formations. The imbricate stack so produced is not observed in the field and both footwall and hangingwall cut-offs must have been eroded during the uplift of the NPDZ.

1) Emergent thrusting and footwall collapse



2) Wedging and monocline roll-back

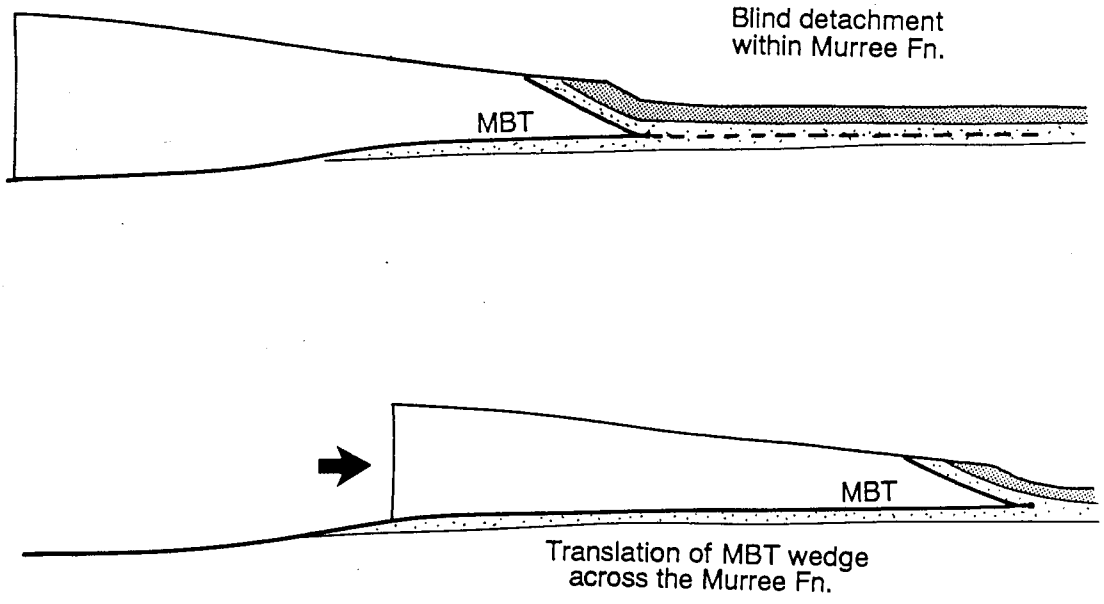


Fig.8.6 CONTRASTING MODELS OF MBT DISPLACEMENT

A further possibility is that the L.Siwaliks were not deposited in this region, the molasse depocentre having migrated south to the northern Potwar by this time (8–6Ma). In this case the MBT would have propagated across the surface of the molasse basin like, for example, the SRT today. The total thickness of the L.Siwaliks at Mahesian is *ca.*500m (Imran Khan, pers.comm.). In the most northerly outcrops of the L.Siwaliks in the Soan Syncline, they attain a thickness of *ca.*1000m (Behrensmeyer, 1987). This suggests that the L.Siwaliks were deposited further north but have since been eroded.

Displacement on the Hazara Detachment (MBT) was *ca.*85–90km. The cause of such large displacement on a single detachment is uncertain. Gravity sliding is often invoked as a mechanism for the translation of major thrust nappes over large distances. The overpressured molasse sediments can provide easy–slip detachment. How is a foreland–dipping slope generated on this detachment? One possible means is the uplift of the internal zones that resulted in the cooling of the metamorphic pile during the period 24–15Ma (Treloar & Rex, 1990). The uplift of this zone produced gravitational instability which caused the extensional reactivation of structures such as the MMT (Williams, 1990). Could this extensional collapse have been southerly–directed also? If the uplift of the internal zones is attributed to isostatic readjustments then the Hazara Detachment could have been southerly–dipping at this time, thereby inducing gravitational sliding. I do not believe MBT displacement was gravity–assisted. Rather, I attribute prolonged translation on this single structure to steady–state, supercritical wedge propagation (Fig.8.7).

8.3.3. Duplex Formation in the Footwall to the MBT.

Subsequent to translation on the MBT, the Hazara–Potwar thrust wedge propagated by collapse of the MBT footwall ramp. This produced a duplex between a floorthrust which lay within the Hazara Group close to the basement–cover contact and a roofthrust which lay within the Murree Formation close to the MBT. The roofthrust may, in fact, have reactivated the MBT at first, but towards the south it probably lies within the molasse beneath the MBT. The

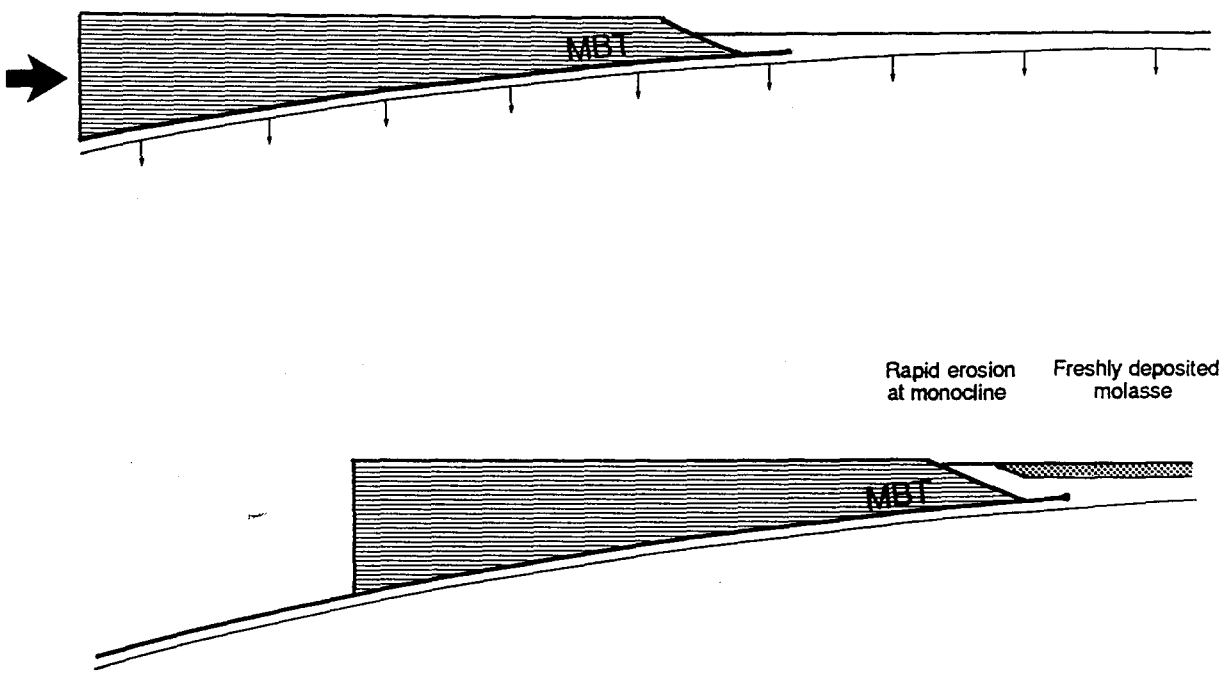


Fig.8.7 CARTOON ILLUSTRATING THE "STEADY-STATE" PROPAGATION OF THE MBT WEDGE

Subsidence of the footwall generates space in the foreland basin for further sediment accumulation during which time the wedge advances along the same decollement horizon. No wedge accretion occurs during propagation as the molasse in front of the wedge is continually uplifted and eroded. Note that the increasing size of the monocline may eventually force a change in propagation style.

most southerly part of the duplex underlies the northern Potwar where it is called the NPDZ. Since the MBT nappe had been translated 85–90km across the molasse basin during the early stages of duplex formation the duplex was overlain by a considerable overburden. In the later stages the duplex propagated beyond the MBT nappe and was overlain by molasse alone. This change in overburden is thought to have caused a change in roofthrust vergence from forethrust to backthrust. Certainly, the latter period of NPDZ formation is associated with backthrusting and backfolding of the overlying strata (Fig.8.8) – such folds being recognised as far north as Bagnotar in the "Slate Zone" of the foothills. The superposition of this backfolding phase onto the earlier imbrication of the MBT hangingwall results in *ca.*65% shortening in this zone.

This duplex underlies the entire region between the Soan Syncline and the northernmost exposures of molasse in the HKS, near Balakot. Estimates of the shortening it accommodates are poorly constrained due to variations in stratigraphic thickness towards the north and uncertainties related to the depth to basement and the depth to the MBT. In the northern Potwar, where the duplex is best constrained, approximately 40% shortening is estimated. The thickening of the Eocambrian–Eocene wedge towards the north means that less shortening is required in that direction to produce equivalent uplift. Therefore perhaps 30–35km shortening is accommodated in the duplex. This shortening occurred between 6–4Ma since it predated movement on the SRT (4.5Ma).

Backfolding associated with the NPDZ is responsible for the change in mountain front style at the MBT between Kashmir and Hazara. In Kashmir, the mountain front is formed by an emergent thrust bringing Eocene carbonates over Miocene molasse (Burbank, Raynolds & Johnson, 1986). In Hazara, because the MBT has been backfolded during the formation of the NPDZ duplex, the mountain front is formed by an anticline. The MBT is not therefore emergent in the Margalla and Murree Hills of Hazara (Izatt, 1990).

Uplift associated with the NPDZ increases towards the west where, in Kohat, the imbricated carbonate succession is thicker. In this area the roofthrust lies within the Eocene

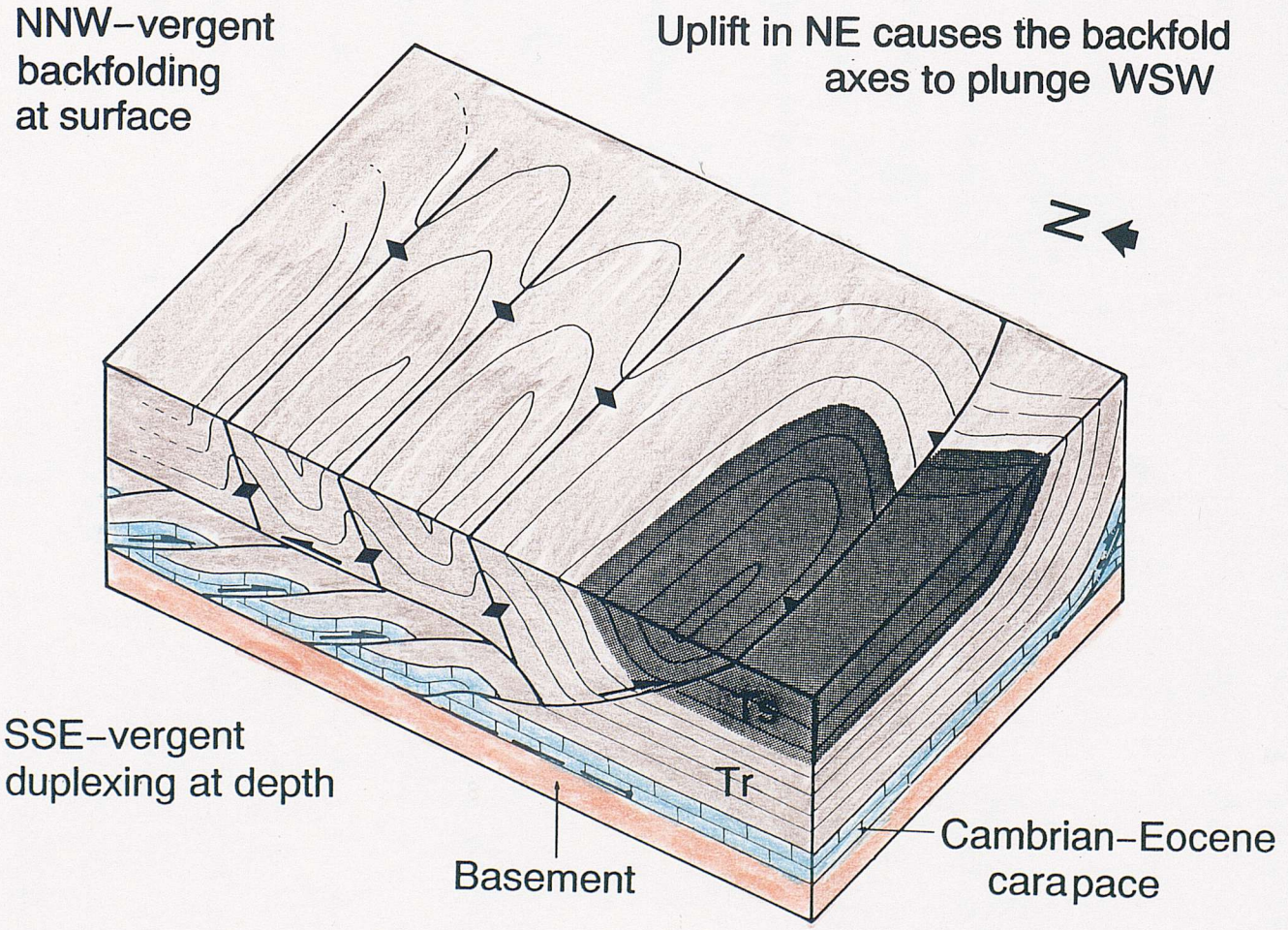


Fig.8.8 CARTOON ILLUSTRATING THE BASIC STRUCTURE OF THE NPDZ.

Panoba shale unit (Izatt, 1990) indicating that the roofthrust cuts down section towards the west. Since the basement dips towards the northeast the ramping of the roofthrust may be an attempt to maintain the same topographic elevation.

The backfolding of the earlier MBT-related structures results in subvertical bedding orientations throughout the Galiat and the Kallachitta Ranges to the west (Izatt, 1990). This deformation produced major uplift and thickening of the wedge. This infers that the critical wedge taper and resistance on the basal detachment were both high at that time. It is argued therefore that NPDZ formation occurred whilst the basal detachment lay within the Hazara Group greywackes. The facies transition between the Eocambrian Hazara Group greywackes and Salt Range Formation evaporites caused a major decrease in resistance on the basal detachment and a major change in wedge propagation style. Once the basal detachment had encountered the Salt Range Formation, NPDZ duplex formation at depth and backfolding at the surface ceased and the deformation front propagated rapidly south beneath the Potwar Plateau.

8.3.4. Displacement on the Salt Range Detachment (SRT).

At approximately 4.5Ma the style of deformation in the Hazara-Potwar thrust wedge changed completely when the basal detachment encountered the Eocambrian Salt Range Formation which comprises halite deposits interbedded with red shales, clays and other evaporites. The salt is over 2000m thick in some anticlinal cores, but averages *ca.*200-300m thickness (Pennock et al, 1989), in the eastern Potwar. It reduces the basal resistance of the wedge and therefore the critical taper. The wedge has a low surface slope and the structures of the southeastern Potwar are generally symmetrical, as is predicted in a wedge that has aweak basal detachment (Davis & Engelder, 1985). These structures develop initially as **salt pillows** that may nucleate above basement offsets, and grow into **pop-ups** which transfer displacement from the SRT to a decollement within the basal Siwalik Group molasse. Further shortening produces **low angle breaching thrusts** that breach the intramolasse decollement and ramp through the Siwalik succession. The various stages of this evolution can be observed along strike

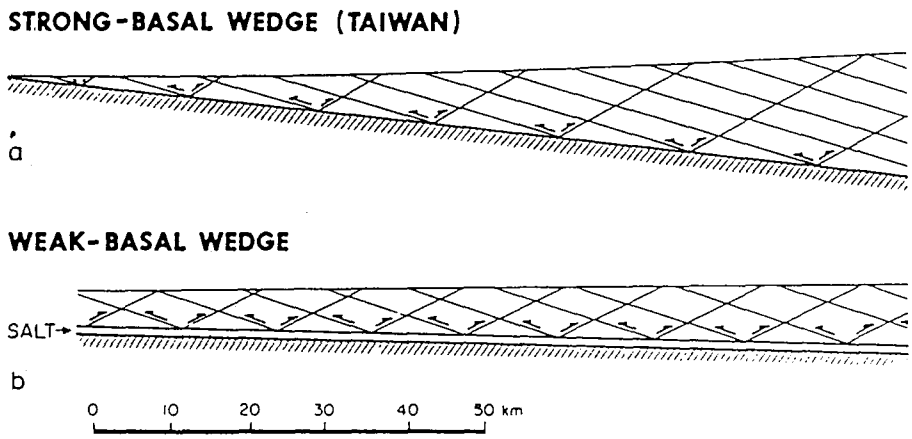


Fig.8.9 CONTRASTING WEDGE CHARACTER AND THE EFFECT OF SALT
(from Davis & Engelder, 1985)

on individual structures, e.g. Domeli.

The association of basement offsets with thrust ramp nucleation was proposed by Wiltschko & Eastman (1983). After restoration the Domeli ramp coincides with an apparent basement offset (Pennock et al, 1989) and ramping of the SRT has been linked to basement normal faulting (Lillie et al, 1986; Yeats & Hussain, 1987). This fault is E–W trending with downthrow towards the north. Seismic imaging of the other apparent offsets is poor and their trend has not been established.

Under the Himalayan compressive stress regime the salt which immediately overlies the Precambrian basement began to flow, forming salt pillows. Ponding of salt at the Salt Range normal fault scarp uplifted the Cambrian–Eocene carapace to produce the emergent the SRT. Initial uplift on the SRT has been dated at 4.5Ma by the presence of pebbles of Tobra Formation of Permian age within the U.Siwalik (Dhok Pathan Fm.) molasse at Bhaun (Baker, 1987).

The flow of salt led to the development of broad synclines and tight, linear anticlines (Yeats & Lillie, 1991). With increasing uplift and shortening the anticline crests were breached by N– and S–vergent, small displacement thrusts which ramped through the Cambrian–Eocene carapace before flattening into an intramolasse detachment. The level of this detachment is constrained to the lower Siwalik or uppermost Rawalpindi Group by a combination of seismic data and field exposures. Exploration drilling in the Potwar has encountered very high fluid pressures within the L.Siwalik units (unpublished O.G.D.C data) and it is assumed that these levels have been exploited as easy–slip detachments. The undulose intramolasse detachment is predicted to dip gently northward from 2–3km below datum in the southern Potwar to 3–4km below datum near Riwat (Fig.8.4). The shortening accommodated at depth by the pop–ups is accommodated by small splays that rise off the detachment at culminations and terminate within the U.Siwaliks as tipfolds. The surface strata are often steeply–dipping in contrast to the gentle dips at depth. This stage of structural evolution is

typified by the Buttar, Ghungrila–Kallar and Qazian structures which accommodate only *ca.*5.5km, 3km and 4.5km shortening respectively. Importantly, they accommodate between 1.5–2km uplift.

The final stage of structural evolution is the breaching of the intramolasse detachment and the formation of a through–going low angle thrust. Two examples of this stage occur in the SE.Potwar – the Domeli Thrust and the Mahesian/Jogi Tilla Thrust. Both these thrusts are emergent and ramp through the molasse at between 8–15°. Pennock et al (1989) estimate 8.8km shortening on the Domeli Thrust but this figure decreases towards the northeast where the surface expression of the structure decreases and the thrust finally terminates as a box–shaped tipfold. The proposed origin of these structures as pop–up bounding thrusts that breach through the intramolasse detachment is supported by the manner in which they change polarity along strike, as displayed by the Mahesian/Jogi Tilla Thrust. The NW–vergent Mahesian Thrust links with the SE–vergent Jogi Tilla Thrust in the vicinity of the G.T. road. The transfer zone trends N–S and has been strongly rotated anticlockwise during displacement.

The Domeli and Dil Jabba Thrusts converge to produce a polarity change along strike. This is produced, however, by the convergence of two independent structures in the same way that the Jogi Tilla and Rohtas structures merge along strike.

The thrust sheets of the SE.Potwar have suffered anticlockwise rotations (Johnson et al, 1979; Opdyke et al, 1982; Burbank & Raynolds, 1988) of 10–60°. Anticlockwise rotations of 56° occurred in the southeastern Salt Range (Treloar et al, in press). The large rotations can be explained in terms of pinning of individual thrusts, e.g. the pinning of the Salt Range Thrust under the Chambal Ridge. The NE–SW trend of most of the structures of the eastern Potwar is the result of large scale, eastward pinning of the Hazara–Potwar thrust wedge during the formation of the HKS. Recently uplifted structures such as Rohtas have not been rotated (Johnson et al, 1979) and this suggests that the folds have developed approximately perpen–dicular to the NNW–SSE transport direction and have been rotated anticlockwise during

subsequent displacement on the SRT. This rotation is also observed in the transport direction indicators of the NPDZ and Galiat regions. Between the Kala Chitta and Murree Hills lineations fan from N210°–220°E to N155°–160°E – a rotation of 50–60°. To what extent this rotation reflects passive rotation during NE–SW compression (which amounts to 12–15% shortening in the Galiat), rotation during differential S–vergent displacement or, indeed, an original transport direction fan is unknown.

Deformation since 4.5Ma in the eastern Potwar has been minor, the structures accommodating only *c.*33km or 18% shortening and little vertical thickening. This is equivalent to a time–averaged shortening rate of 7mm/yr – one sixth of the rate during MBT translation. Palaeomagnetic polarity reversal dating has been used to try to date the onset of uplift on the structures of the eastern Potwar (Johnson et al, 1979; 1982; 1986; Opdyke et al, 1979). This data constrains the deformation of the southeastern Potwar to the period 3–1.6Ma, during which time the sediment accumulation rate fell. It does not elucidate the sequence of deformation. However a general southward propagation of the deformation front can be observed since the Chambal Ridge attained surface expression <0.7Ma ago (Opdyke et al, 1979) and the Rohtas and Pabbi Hills structures were expressed only 0.4Ma ago (Johnson et al, 1979).

8.4. VARIATIONS IN STRUCTURAL STYLE ACROSS THE POTWAR PLATEAU.

During the formation of the numerous fold and thrust structures of the eastern Potwar since 4.5Ma, displacement in the western and central Potwar has been accommodated on a single structure, the emergent SRT (Leathers, 1987; Baker, 1987). Why did this change in structural style along strike in the Potwar occur? Lillie et al (1987) proposed that the topographic slope of the wedge had to increase towards the east in order to remain supercritical. He suggested two reasons for this.

a) A decrease in the basal slope of the wedge. Since the topographic slope of the Potwar is negligible, any decrease in the basal slope of the wedge will decrease the wedge taper. Seismic data from the western and eastern Potwar suggests that the northerly basement dip

decreases from west to east across the Potwar from 2–3° to <1° (Jaume & Lillie, 1988), although the top basement reflector is poorly imaged in the eastern Potwar.

b) An increase in the basal friction of the wedge. Lillie et al (1987) suggest that the margin of the salt basin is approached towards the east and therefore the salt decollement becomes less effective. Salt thicknesses of over 2000m have been drilled within the cores of pop–ups in the eastern Potwar, suggesting that average salt thickness still exceeds 200m in this area (Pennock et al, 1989). Butler et al (1987) propose that the SRT ramps into the molasse series towards the east.

Davis, Suppe & Dahlen (1983) describe three factors that control the propagation of a wedge on its basal detachment – the wedge taper, the basal friction and the internal strength of the wedge. It is my view that this last factor dominates structural style variations along strike in the Potwar. Field mapping (Gee, 1981) and well log data (S.M. Naeem, *pers.comm.*) show that the stratigraphic template of the undeformed Potwar varied markedly from west to east across the Potwar (Fig.6.8). In rheological terms the stratigraphy of the Potwar can be divided into three units:–

a) the Infracambrian salt,

b) the Cambrian–Eocene shelf carapace and

c) the Miocene–Recent molasse.

The first unit is uniform throughout the Potwar and provides the basal detachment upon which the upper two units are translated. The second unit is a competent, brittle unit comprising predominantly Gondwanian sediments and post–Gondwanian break–up passive margin sediments. The third unit comprises a thick, incompetent and ductile series of poorly–consolidated shales and sands. Where present the Cambrian–Eocene carapace dictates a structural style typified by the western and central Potwar, ie. major thrust sheets that are translated over the salt. In the absence of this carapace, or where it becomes very thin, the molasse series controls structural style and this occurs in the eastern Potwar.

The controlling factor of structural style in the Potwar is therefore the thickness of the Cambrian–Eocene carapace. In the western Potwar 1180m of carapace sediments are present in the Dhermund well (Leathers, 1987). In the eastern Potwar the carapace has thinned to 400m in Qazian well (S.M. Naeem, pers.comm.) and this figure reduces further towards the extreme southeastern Potwar (240m at Warnali). This reduction is partially a product of stratigraphic thinning of the Mesozoic passive margin wedge onto the Indian craton. **The southeastward thinning of the Cambrian–Cretaceous sequence is due to erosion at the base Palaeocene unconformity** (Figs.6.8 & 6.11).

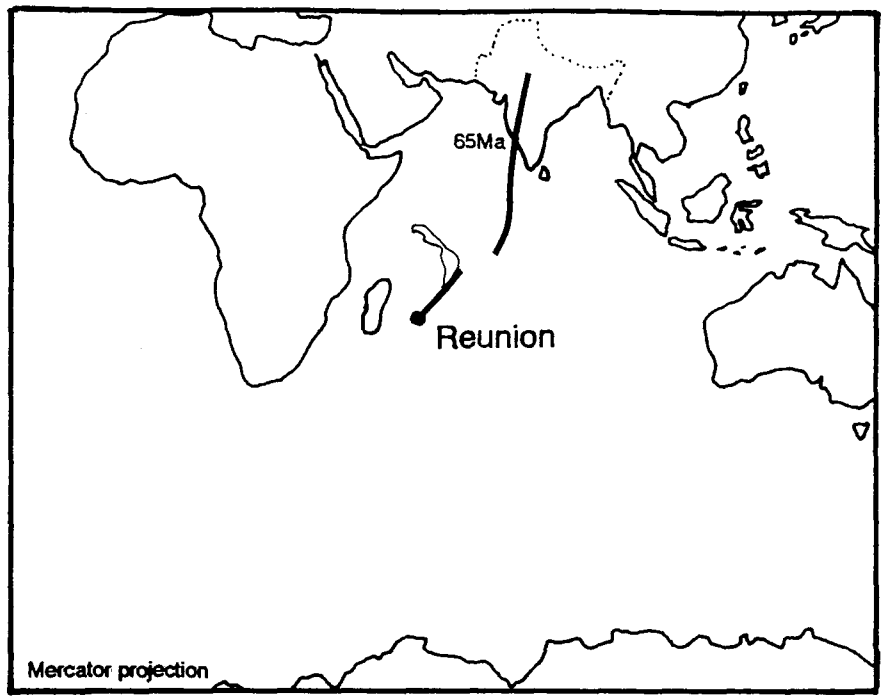
The base Palaeocene unconformity occurs throughout the western Indian subcontinent and marks a period of subaerial erosion. The unconformity is represented by the Hangu Formation in Hazara and the Potwar/Salt Range region and correlates with the Stumpata conglomerate of the Zanskar shelf (Garzanti, Baud & Mascle, 1987). This indicates that subaerial conditions reigned over a SSW–NNE trending swath across Indian margin, extending to the shelf–slope break itself. The unconformity is also observed towards the west, the Cretaceous–Tertiary boundary being marked by a disconformity between the Pab and Dunghan Formations in the Sulaiman Ranges (Hemphill & Kidwai, 1973). At the Sanjawi Arch, near Quetta, Palaeocene strata overlie middle Cretaceous strata whilst further south, at Muslimbagh and the Kirthar Ranges, Eocene strata overlie Jurassic rocks (Shah, 1977, Izatt, *pers.comm.*).

The origin of this unconformity is uncertain but it represents a major regressive–transgressive cycle that exposed the shelf–slope break at a time when world sea–levels were relatively higher than today (Haq et al, 1983). At the time (65Ma), Eurasia and India were still a long way from continent–continent collision, the Neotethys Ocean being over 2000km wide.

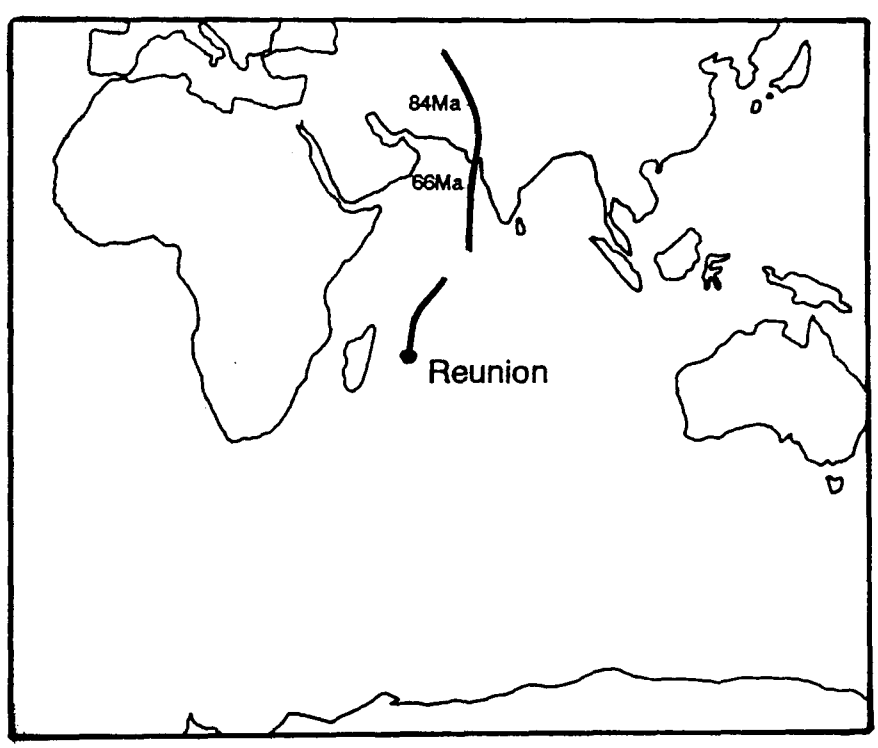
The unconformity is commonly attributed to ophiolite obduction onto the northern margin of the Indian plate (Searle, 1983; 1986). That ophiolite obduction is associated with the foundering of the continental margin (Searle, 1983) rather than its emergence would argue against an obduction–related origin for the base Palaeocene unconformity. In addition, numerous

authors have proposed that obduction occurred in the Eocene (Frank, Gansser & Trommsdorff, 1977; Thakur, 1981; Garzanti, Baud & Mascle, 1987).

The end Cretaceous–early Palaeocene period is marked by a major fall in relative sea–level and here I suggest that it was caused by the drift of India across the Reunion hotspot (Fig.8.10). Plate reconstructions (Johnson, Powell & Veevers, 1976; Norton & Sclater, 1979; Barron, 1981; Morgan, 1981; Besse & Courtillot, 1988; Scotese, Gahagan & Larson, 1988; Dewey, Cande & Pitman, 1989) are in agreement that in the late Cretaceous (between 85–65Ma) India drifted rapidly northwards across the present position of the Reunion hotspot. Latitudinal movement is well–constrained, longitudinal movement less so. At 65Ma the Deccan Traps were extruded during a very short period (<1Ma) that is related to the rifting of the Seychelles from India. The association of flood basalts and continental rifting has been linked to uplift, extension and decompressive melting due to thermal anomalies, or hotspots, within the mantle (White & McKenzie, 1989). Such anomalies, which typically reach 100–200°C, extend across a 1000–2000km diameter region and produce uplift of up to 1000–2000m. Simple isostatic calculations suggest that a 100°C thermal anomaly would produce between 600–800m uplift whilst a 200°C anomaly would produce 1200–1800m uplift. The rapid northward drift of India may have prevented full development of the uplifted dome. Even the minimum estimate could generate uplift sufficient to expose the entire continental shelf of northern India. It is proposed that this happened at the end of the Cretaceous causing emergence of the entire shelf and widespread laterite formation. Once northern India had passed over the centre of the anomaly the continent began to thermally subside leading to marine transgression across the partially eroded shelf in the Lower Palaeocene. In the Deccan to the south, uplift above the thermal anomaly caused rifting and decompressive magmatism. Only in this region, where the melt was added to the crust and the lithospheric mantle was depleted, did uplift persist to this day. The uplifted dome has a diameter of approximately 2000km (Cox, 1989; White & McKenzie, 1989).



1) by MORGAN (1981)



2) by SCOTESE, GAHAGAN & LARSON (1988)

Fig.8.10 PREDICTED TRAIL OF THE REUNION HOTSPOT

8.5. THRUST WEDGE INTERFERENCE AND THE FORMATION OF THE HKS.

During the last 4.5Ma the western limb of the HKS has also been subjected to compression normal to the propagation direction of the Hazara–Potwar thrust wedge. This is associated with the advance of the Main Himalayan thrust wedge towards the southwest. Several structures belonging to this system have been dated (Burbank, 1982; Burbank, Raynolds & Johnson, 1986) such as the MBT in Kashmir (3Ma), the Kotli Thrust (2.8–1.5Ma) and the Suruh anticline (<1.4Ma). Clearly, therefore, two directions of tectonic transport were active concurrently during the deformation of the Potwar.

At 4.5Ma (Baker, 1987) the SRT initially achieved surface expression. This represents a phase of rapid propagation of the deformation tip of the SSE–vergent, Hazara–Potwar thrust wedge due to the presence of a salt decollement. At this time, the SW–vergent Kashmiri thrust wedge had not encountered this salt horizon and was therefore propagating more slowly. Uplift on the MBT in Kashmir did not occur until *ca.* 4Ma when ponding of the Kashmir Basin commenced (Burbank, 1983). Hence, since that time the Kashmiri thrust wedge has advanced southwestward into the Hazara–Potwar thrust wedge. Instead of advancing into the undeformed Himalayan molasse basin, the Kashmiri thrust wedge has encountered an already thickened and deformed cover sequence (Fig.8.11).

The hindrance to propagation becomes more accentuated towards the north where the Hazara–Potwar wedge is over 12km thick. This effect is well–illustrated by the Kotli Thrust which is a SW–vergent structure lying in the MBT footwall in Jammu and Kashmir. In Jammu, the Kotli thrust carries Eocambrian Dogra Slates over Miocene molasse (Ashraf, Chaudhry & Qureishi, 1983) and hence, the sole thrust of the Kashmiri wedge lies close to the basement–cover contact within the Dogra Slates at this point (the Dogra and Hazara Slates are lateral equivalents (Wadia, 1928)). The facies transition between Hazara/Dogra Slates and the Salt Range evaporites therefore lies to the southwest of Kotli. In Jammu, the Kotli thrust lies to the northeast of the limit of SSE–vergent Potwar deformation. As one traces the Kotli Thrust

northwestwards along strike it converges with the lateral limit of the Hazara–Potwar thrust wedge and to the southeast of Murree the two thrust wedges interfere. This is expressed by the uplift of eastern Hazara (Treloar et al, in press) (Fig.8.11).

The geometry of the SW–vergent structures in the footwall of the Kashmiri MBT is poorly constrained due to the absence of seismic coverage in this region and the restricted access for foreigners for security reasons to the area. I propose that the Kashmiri thrust wedge is currently wedging beneath the Hazara–Potwar thrust wedge and that the latter is uplifted above the frontal triangle zone of the former (Fig.8.12).

This is thought to have occurred due to the gross geometry of the foreland basin in the NW Himalayas. Since the foreland basin has formed in response to loading by the Hazara–Potwar and Kashmiri thrust wedges it descends towards both the northeast and north–northwest. It is thought therefore that the basal detachment of the Hazara–Potwar wedge ramps towards the northeast (Butler et al, 1987) into the molasse sequence in order to maintain its topographic level. Thus, as the SW–vergent Kashmir thrust wedge advanced on a sole thrust within the Dogra Slates, it imbricated the Eocambrian–Eocene strata in the footwall to the Hazara–Potwar wedge.

The Kotli Thrust and the blind thrusts in its footwall which detach the Dogra Slates, have propagated by wedging into the eastern flank of the Hazara–Potwar wedge. It is proposed that the SRT detachment of the Hazara–Potwar wedge has been reactivated as a frontal backthrust to the Kashmiri thrust wedge. The large scale frontal monocline related to this triangle zone is reflected in the W–plunging folds of the Galiat and the eastern NPDZ. It is probable that the imbricates of the NPDZ duplex have been reactivated to some extent.

The overburden of the Hazara–Potwar thrust wedge has pinned the propagation of the Kashmiri wedge resulting in thickening of the Kashmiri wedge. During this process the overlying Hazara–Potwar wedge has been uplifted and overturned. The Galiat remains topographically elevated today. In this area, the superposition of SW–vergent compression upon

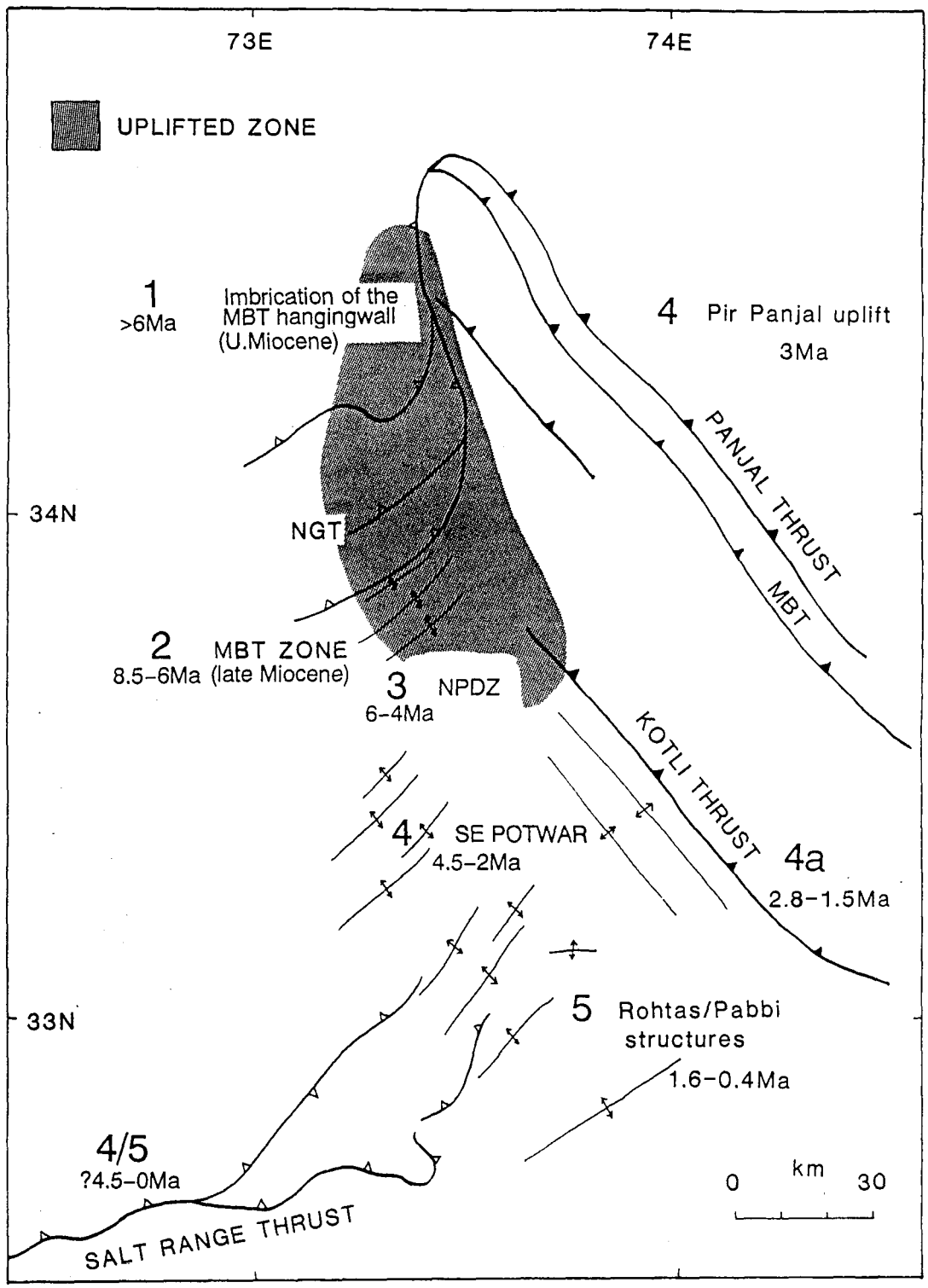
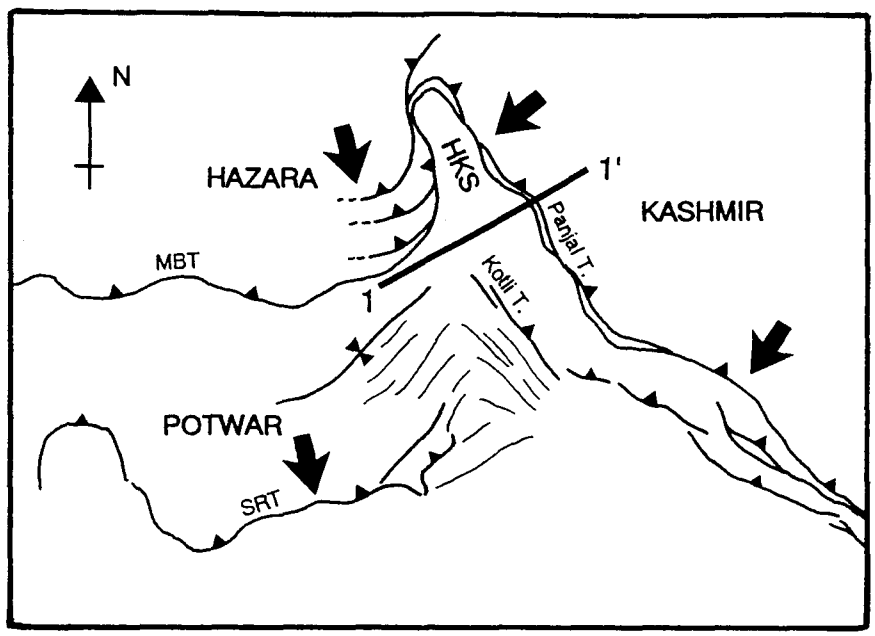


Fig.8.11 THRUST INTERFERENCE AND TIMING IN THE HAZARA-KASHMIR SYNTAXIS

Deformation Sequence, 1-5



Arrows indicate tectonic transport direction.
Section 1–1' is drawn parallel to the Kashmiri transport direction, but perpendicular to the Hazara–Potwar transport direction.

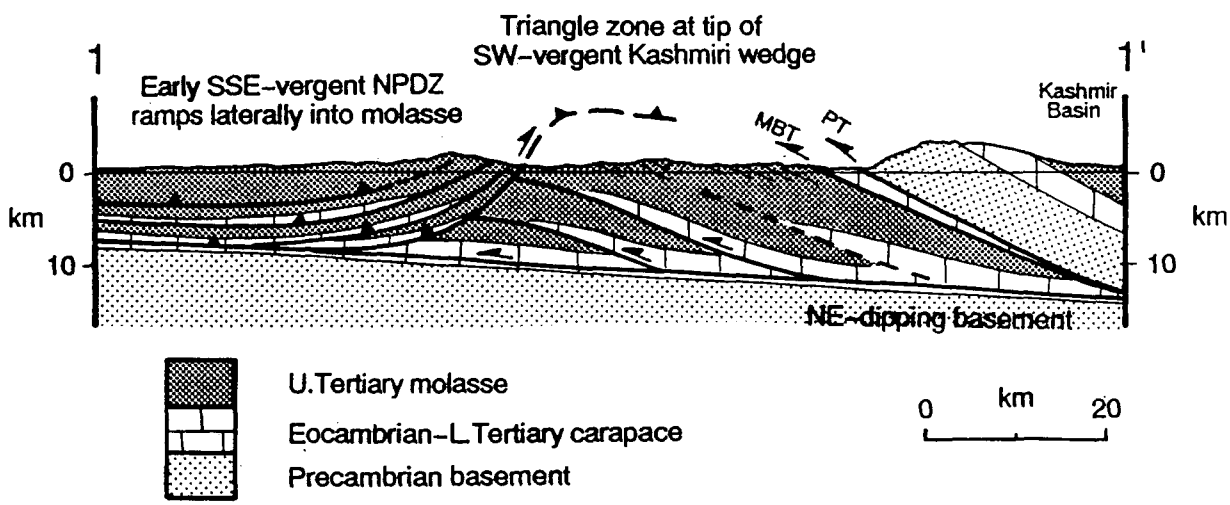


Fig.8.12 STRUCTURE OF THE HAZARA–KASHMIR SYNTAXIS, showing wedging of the Kashmiri wedge beneath the Hazara–Potwar wedge.

the early SSE–vergent structures has resulted in **crossfolding** about NNW–SSE trending axes.

Hence, the southwestward propagation of the Kashmiri thrust wedge has monoclinally–folded and uplifted the strata in the hangingwall to the Hazara Detachment and also produced approximately 12–15% shortening in front of the monocline by buckling.

Both of these effects would cause local anticlockwise rotation of lineation data. In the buckled zone, a 10% NE–SW shortening would produce 4–6° anticlockwise passive rotation of an originally SSE–directed lineation, assuming pure shear. Rotations associated with monoclinal folding about a NW–SE trending axis are of the same order but restricted to a smaller area. Neither mechanism can therefore explain the fanning of transport direction indicators from N210°E in the western Kalachitta Hills to N155°E in the Galiat and eastern NPDZ.

This fanning is attributed to eastward pinning of the Hazara–Potwar thrust wedge by the advancing Kashmiri thrust wedge. Anticlockwise rotations have been recorded in the Hill Ranges (55°; Izatt, 1990 and this work), the Soan Syncline (39°; Burbank, 1984), the southeastern Potwar (24–31°; Opdyke et al, 1982). The rotation decreases towards the foreland suggesting that the rotation is cumulative and therefore the Hazara–Potwar thrust wedge has been pinned in the east since at least 6Ma.

On the eastern limb of the syntaxis the Kashmir thrust wedge appears to be pinned in the northwest by the Hazara–Potwar thrust wedge. Klootwijk et al (1986) report 22–39° clockwise rotations in Jammu, whilst closer to the proposed pinning point in Kashmir, Bossart et al (1990) measured a rotation of 43°.

The two convergent thrust wedges have both experienced strong rotations due to lateral pinning and interference during propagation. This situation is occurring actively today but is predicted to be a feature of the NW Syntaxis throughout its evolution since collision. Ultimately, such interference is a product of the shape of Greater India as it subducted under the

Asian plate margin.

CHAPTER 9

CONCLUSIONS

9.1. The External French Alps.

The principal conclusion of fieldwork to this thesis in the environs of the Pelvoux massif in the external French Alps is that progressive Alpine compression can be divided into three distinct phases. These comprise :

a) The Pre-Nummulitic phase of the Arvinche.

b) The Pre-Nummulitic phase of Pelvoux.

c) The Post-Nummulitic phase.

In the literature, the Pre-Nummulitic has been described as a single N-vergent event in both Pelvoux (Gidon, 1979) and the Arvinche (Barbier & Barfety, 1972). Alternatively it has been dismissed as a minor event that was restricted to Pelvoux alone (Beach, 1981). The new field data presented in Chapter 4 indicates that the Pre-Nummulitic events of the Arvinche and Pelvoux regions can be distinguished on the basis of structural style despite the absence of other dating criteria.

9.1.1. The Pre-Nummulitic phase of the Arvinche.

a) This event occurred between the U.Jurassic and U.Eocene. It is tentatively related to the Pyrénéan-Provençal phase of Alpine shortening, but may be of Pre-Senonian age.

b) The phase is characterised by N-vergent, asymmetric folding of the U.Jurassic sequence.

c) At Triassic-L.Jurassic horizons the phase is characterised by evaporite-cored anticlines that formed by the amplification of diapirs. Original diapirism is thought to be of Toarcian-Aalenian age (Mascle et al, 1988; Samec et al, 1988), contemporaneous with major rifting.

d) Disharmony is accommodated by the U.Lias–Dogger shale sequence.

e) N–vergent thrusting of basement and Triassic strata in the Grandes Rousses and Rochail basement fault blocks (Barfety & Gidon, 1990) may have occurred at this time.

9.1.2. The Pre–Nummulitic phase of Pelvoux.

a) This phase is also constrained to the U.Jurassic–U.Eocene period, but is thought to postdate the Pyrénéan–Provençal phase.

b) The tectonic transport direction was predominantly WSW–vergent.

c) Three styles of deformation are recognised:

i) inversion geometries,

ii) large scale fault–propagation folds involving basement and cover rocks (preserved as footwall synclines) and

iii) thrust imbrication of the Triassic strata.

d) The strike–slip mosaic model of Gidon (1979) is not supported by fieldwork to this thesis. A thick–skinned inversion model, proposing the selective reactivation of inherited Mesozoic faults, better explains the observed field relationships.

e) A major effect of this event was 1–2km uplift that caused the exposure of Hercynian basement prior to the Nummulitic marine transgression.

f) This phase may be expressed in the southern part of the Ultradauphinois Zone by imbrication of the dolomitic Triassic sequence. This is denied by Beach (1981), but provides a lateral termination of Pre–Nummulitic displacement on the Combeynot thrust.

9.1.3. The Post–Nummulitic phase.

a) This phase is recognised in southern and eastern Pelvoux and the Arvinche by the involvement

of the Nummulitic flysch series, but it is postulated that it also exists in central and northern Pelvoux where the flysch has since been eroded.

b) Although backfolding (Platt et al, 1989) and extensional faulting (Coward et al, 1991) occurred during this phase, the phase is characterised by WNW- to WSW-vergent compressional deformation. In southern Pelvoux tectonic transport is towards N255°-265°E, but in the Arvinche the tectonic transport direction is N280°E. This divergence may reflect the influence of body forces upon the interplate motion vector (Platt et al, 1989). Local anomalies occur, such as the NW-vergent shearing of the Gres de Champsaur flysch, which are attributed to local stresses associated with uplifted basement blocks.

c) Two major Post-Nummulitic thrusts are recognised:

i) the Frontal Pennine Thrust and

ii) the Ultradauphinois/La Meije/l'Olan/Sirac Thrust system.

d) Both of these systems breached the Nummulitic unconformity, preserving the flysch series in their footwalls. Deformation intensity decreases as one moves westward from each of these major thrust traces (Tricart, 1984).

e) Imbrication of the Mesozoic sediments of the southern Ultradauphinois Zone is of this age (Beach, 1981), although Pre-Nummulitic shortening is not precluded.

f) Displacement on the Sirac Thrust has juxtaposed two U.Eocene flysch series of different provenance. Shortening on the structure may therefore greatly exceed the minimum value of 12km dictated by cut-off relationships.

g) A minimum of 2-3km Post-Nummulitic uplift is recorded by the erosion of the Nummulitic flysch series from central Pelvoux and the tilting of the basal Nummulitic unconformity. Similar uplift was experienced in the Arvinche.

9.2 The External Himalayas of Pakistan.

The primary conclusion of the fieldwork to this thesis carried out in the Pakistani Himalayas is that **the Hazara–Kashmir Syntaxis evolved as a result of interference between two concurrent, but convergent, thrust wedges – the SSE–vergent Hazara/Potwar thrust wedge and the SW–vergent Kashmiri thrust wedge.** The Hazara–Kashmir Syntaxis is simply the most recent expression of a zone of thrust interference (the NW Syntaxis) that has evolved since initial Himalayan collision. The thrust wedges are convergent due to the shape of the continental margin of the subducting Indian plate at collision, to the action of gravitational body forces and to the lateral pinning of the Hazara/Potwar and Kashmiri thrust wedges at the thickened syntaxial zone.

The evolution of the western limb of the Hazara–Kashmir Syntaxis is characterised by the following:

a) Two major detachments are recognised – the **Hazara Detachment** (which has been equated with the Main Boundary Thrust of India) and the **Potwar–Salt Range Detachment** (or Salt Range Thrust).

b) SSE–vergent imbrication of the Lesser Himalayan Zone (the Galiat and Hill Ranges) occurred prior to major displacememt on the Hazara Detachment.

c) **A minimum of 80km displacement towards the south–southeast on the Hazara Detachment** is inferred from cut–off relationships. Murree Formation molasse is exposed in the footwall of the Hazara Detachment along the entire length of the western limb of the Hazara–Kashmir Syntaxis, being juxtaposed against Cambrian–Eocene age hangingwall sediments. This displacement occurred in post–Chinji Formation times (Izatt, 1990) and yet no Chinji Formation sediments are preserved in the footwall of the Hazara Detachment. It is concluded therefore that the Hazara Detachment propagated by roll–back of a frontal monocline.

d) Displacement on the Potwar–Salt Range Detachment initiated by collapse of the footwall to the Hazara Detachment, producing an **active roof duplex**. The duplex comprises imbricates of the Cambrian–L.Miocene sediments. Displacement was transferred from the floorthrust (the Potwar–Salt Range Detachment) to the roofthrust (the Hazara Detachment).

e) The duplex propagated beyond the southern limit of the Hazara Detachment, at which point the duplex roofthrust became passive. At depth, therefore, the Cambrian–L.Miocene sedments were imbricated by SSE–vergent thrusts. At the surface, however, the displacement in the hang–ingwall of the duplex was accommodated by NNW–vergent backfolding. This is observed in the Miocene molasse of the North Potwar Deformation Zone and the Cambrian–Miocene sediments of the Hill Ranges. The triangle zone at the front of this **passive roof duplex** outcrops along the northern limb of the Soan Syncline.

f) **In early Pliocene times (*ca.*5–4Ma) the Potwar–Salt Range Detachment encountered the evaporitic Salt Range Formation** – this formation is interpreted as a lateral equivalent of the Hazara Slate Formation. This marked the end of duplexing in the NPDZ, although minor out–of–sequence displacement did occur, and the rapid propagation of the deformation tip to the Salt Range.

g) **The Potwar–Salt Range Detachment becomes emergent as the Salt Range Thrust (SRT)** due to the action of a basement offset.

h) Continued displacement on the SRT in the western and central Salt Ranges is accommodated by out–of–sequence shortening on a number of structures in the eastern Potwar PLateau. These structures evolced in three phases:

1) **Salt pillow development** by halokinesis.

2) **Pop–up formation**. Symmetrical fore– and back–thrusts ramp from the Potwar–Salt Range Detachment through the Cambrian–L.Miocene carapace before flattening onto an **intra–molasse detachment within the Chinji Formation**. This detachment is responsible for the

disharmony between surface and seismic data in the eastern Potwar.

3) **Through-going, low angle thrusting** on either the fore- or back-thrust of the original pop-up structure. On certain structures the transfer of displacement from back- to fore-thrust can be observed along strike.

i) Development of the intramolasse detachment occurred due to high fluid overpressures within the rapidly-deposited molasse series.

j) **The contrast between the deformation styles of the western and eastern Potwar Plateau is attributed to the southeastward truncation of the predominantly carbonate carapace beneath the basal Palaeocene unconformity**. Changes in basement dip and facies changes at the margin of the Eocambrian salt basin may also contribute to the change in structural style.

k) The basal Palaeocene unconformity is observed throughout northwestern India and Pakistan and is attributed to uplift associated with passage of the Indian plate over the Reunion thermal anomaly.

l) The structures described in paragraphs a)-e) have been superimposed by crossfolds that trend NW-SE and resulted from SW-vergent compression. They are the expression of the deformation tip of the SW-vergent Kashmiri thrust wedge. This deformation is of L.Pliocene age since they are observed in the Hill Ranges but do not affect the U.PLiocene-Recent structures of the S.E. Potwar.

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