Tectonics of Sedimentary Basins

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Sedimentation and Tectonics of Fossil Rifts

2

A.M.C. Şengör

Introduction

Rifts are fault-bounded elongate troughs, under or near which the entire thickness of the lithosphere has been reduced during their formation (Fig. 2.1; cf., Şengör and Burke, 1978, p. 419). They form in most tectonic settings (Fig. 2.2) and at *all* stages of the Wilson Cycle of ocean opening and closing (Burke, 1978). Because of their morphology, they are commonly convenient sediment receptacles preserving, in various states of completeness, a record of the tectonic environment in which they originate and/or evolve (see Chapter 3). This record may be hidden owing to burial under younger sedimentary (Fig. 2.3A) and/or tectonic units (Fig. 2.3B), but it is rarely completely destroyed. Associated igneous activity is also common, whose products enrich rifts' archive of geological history. Rift-related metamorphism is more modest than that in orogenic belts, the most extreme cases being known from the "metamorphic core complexes" of the southwestern United States (Armstrong, 1982). Owing to their very widespread occurrence, great versatility in terms of the tectonic environment they inhabit, and the geological documentation they preserve, rifts form an ideal object of study for the historian of the earth.

Many rifts do not survive as rifts. Commonly, when the extension factor (ß = extended width/unextended width: cf., McKenzie, 1978a) grows beyond 3 (cf., Le Pichon and Sibuet, 1981), sea-floor spreading starts opening an ocean and destroys the rift. Remnants of most rifts forming continental margins later fall prey to orogeny (i.e., convergent plate phenomena) and become deformed and metamorphosed beyond easy recognition. Some rifts, however, do not become oceans and terminate their tectonic life during the rift stage and get incorporated into the cemetery of fossil structures of our planet as rifts, comparable to an individual who dies at infancy. It is these that are customarily called "failed rifts" in the geological literature, implying that they "failed to generate an ocean." What are generally called "failed rifts" are, in reality, perfectly successful rifts, *as rifts*, but are "failed oceans" and that is why "failed rift" is an inappropriate designation. Thus "fossil rifts" are preferred in the title of this chapter and not "failed rifts".

Before we begin discussing fossil rifts, it is useful to obtain an unambiguous nomenclature and a practical classification of rifts lest we get lost in the richness of the variability of rifts.

Rift, Graben, and Taphrogen

Although "rift" and "graben" commonly are used interchangeably in the geological literature, it is proposed here to confine "graben" to those structures that do not penetrate the lithosphere (i.e., "thin-skinned") and apply rift to those that do (i.e., "thick-skinned"). This proposition is supported by the history of these two terms.

Graben is a German word meaning a ditch or trench. It entered the language of geology early, via mining. In the miners' jargon, "grabens... are depressions or troughs in horizontal beds, which are much longer than they are wide" (Jacobsson, 1781, cited in Rosenfeld and Schickor, 1969). It did not become a common term, however, until Suess' (1883, p. 166) usage for strips of country subsided between two normal faults. The way Suess used it, especially in relation to the East African Rift Valley, "graben" is equivalent to rift.

For Suess' graben, Gregory (1894, p. 295) used the word "rift valley." "Rift" comes from the root "reve," meaning to tear apart or to pull asunder. Thus, whereas the word "graben" is purely descriptive, "rift" involves an interpretation (i.e., extensional rupturing

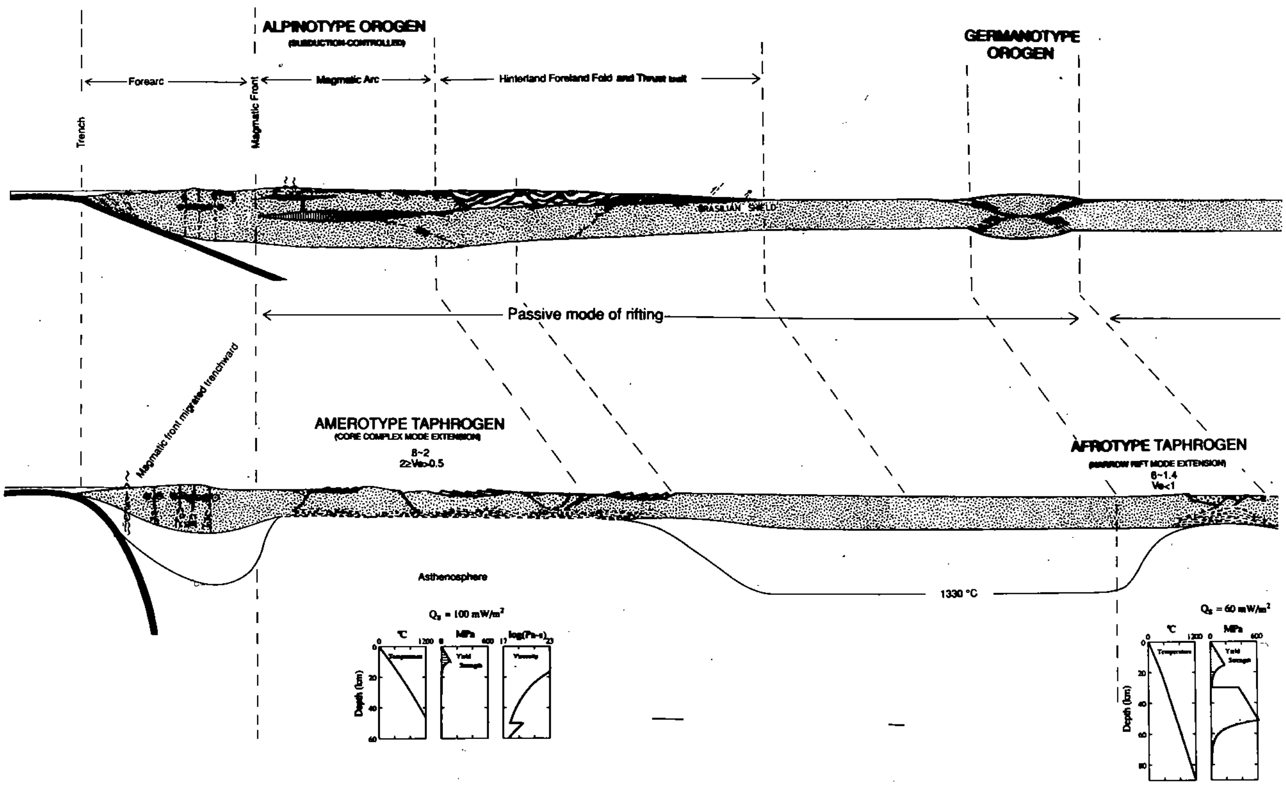
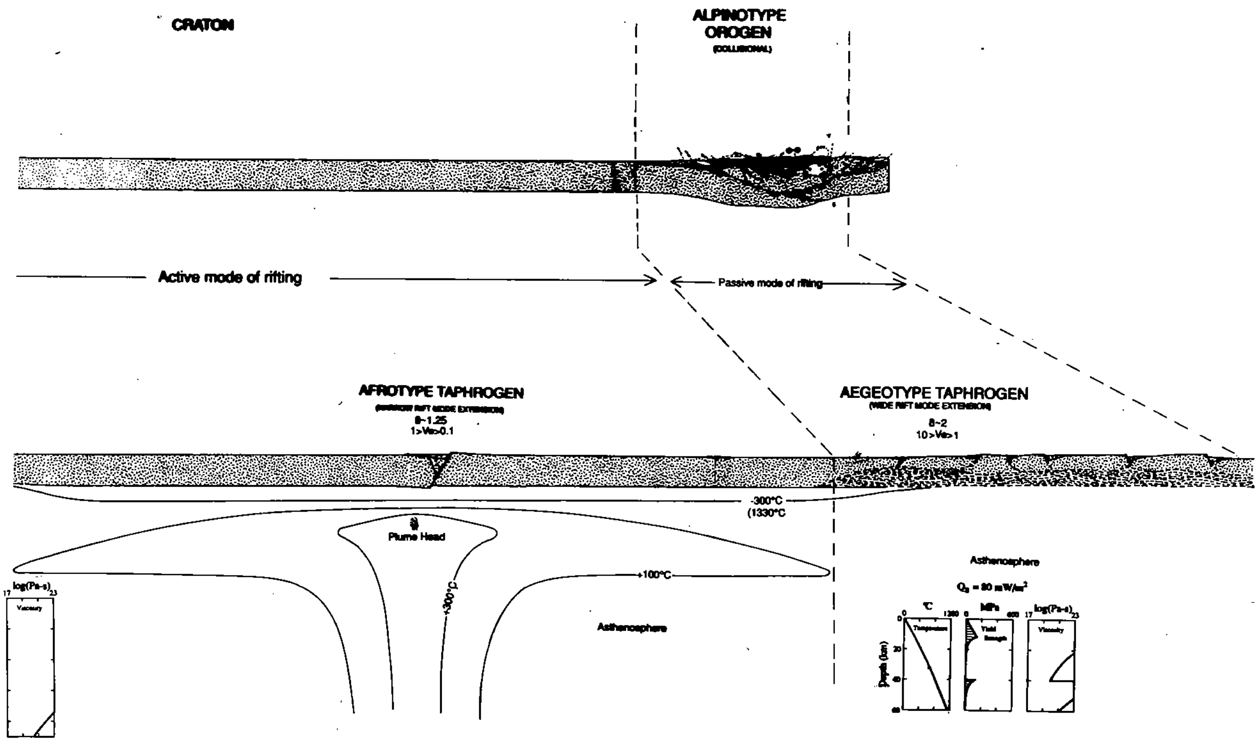


Figure 2.1 Schematic, simplified cross section illustrating concepts of *rift, graben* and *taphrogen*. A hypothetical continent shown in the cross section above, including two *orogenic complexes* (sensu Şengör, 1990a) one of them with both *alpinotype* and *germanotype* parts, located on the margins of a *craton* (cross section at top), is stretched, creating various types of taphrogens including *afrotype*, *amerotype* and *aegeotypes* (cross section at bottom). Notice that quasi-cratonic area, deformed in a germanotype manner during orogeny, leads to an afrotype, largely symmetric rift, immediately under which lithospheric thinning has occurred. By contrast, alpinotype deformed cordilleran-type orogenic area leads to largely asymmetric amerotype rifting, where sub-detachment lithospheric thinning and crustal thinning do not overlie one another at every locality. A similar picture, but this time creating an aegeotype taphrogen, is seen to result from the rifting of a collisional alpinotype orogen (e.g., Aegean Sea rifted atop orogen connecting Hellenides with Turkish ranges). Grabens do not penetrate the lithosphere and form only in areas of superficial deformation, commonly along external margins of taphrogens. For the structural sections, I have consulted papers that reported data on the shallow and deep structure of rifts and grabens, and especially shallow and deep seismic-reflection data. For the amerotype

of a formerly continuous medium). As originally used by the miners, "grabens" referred to smaller and undoubtedly shallower structures than what are called "rifts". It would avoid confusion if one adhered to this distinction in the study of fossil extensional structures. In this chapter, we are concerned almost exclusively with rifts, as defined herein.

Currently, no term is employed to designate *regions of intense extension,* in which many rifts and grabens occur as a result of general lithospheric stretching. For comparable *regions of intense shortening* the term *orogen* has been employed since Kober's (1921, p. 21) suggestion. A corresponding term for zones of intense extension is clearly needed, since some have used the confusing appellation "extending orogens" (e.g., Wernicke, 1981, 1985). *Taphrogen,* derived from Krenkel's (1922, p. 181 and footnote) term *taphrogeny* meaning trough-building (from the Greek τάφροσ = ditch or trench), is suggested to designate the extensional counterparts of orogens. Thus, taphrogens are lithospheric-scale structures, commonly formed from a linked system of rifts and grabens that stretch the lithosphere. Advanced taphrogeny eventually leads to ocean formation (that may be called *thalassogeny,* Kober, 1921, p. 48: from the Greek θάλασσα = sea). If taphrogeny stops before producing ocean (i.e., before leading to thalassogeny), then it leads to subsidence and creates large basins overlying taphrogens (cf., McKenzie, 1978a). In other



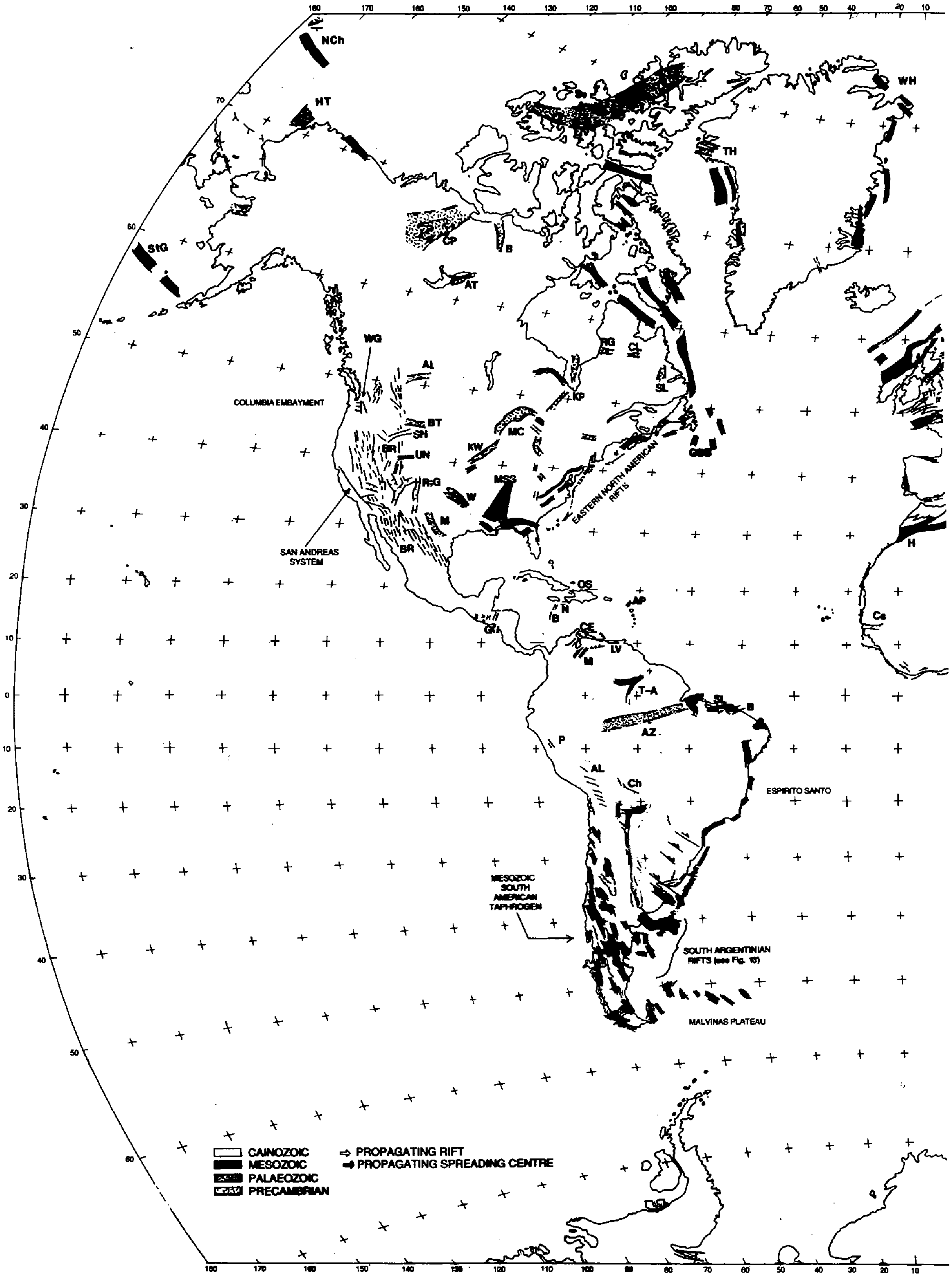
model, I have mainly consulted Allmendinger et al. (1983b, 1986), Miller et al. (1983), Gans et al. (1985), Wernicke (1985), Cheadle et al. (1986) and Hamilton (1987). For the afrotype model, I have used Illies (1970), Rosendahl (1987), and Burgess et al. (1988). For the aegeotype model I utilized Şengör (1982, 1987c) and Eyidoğan and Jackson (1985), plus unpublished seismic-reflection lines in the southern and central Aegean made available to me by courtesy of the late Mr. Ozan Sungurlu of the Turkish Petroleum Co. Owing to the small scale of the diagram such details of normal faulting geometries as those reported by Colletta and Angelier (1982) or Miller et al. (1983) could not be shown. The three graphs for each type of taphrogen at the bottom of the figure show initial model geotherms, yield strengths (for a strain rate of $8 \times 10^{-15}\ s^{-1}$), effective viscosities (for dry quartz crust overlying a dry olivine mantle) for different crustal thicknesses (50km for amerotype, 40km for aegeotype and 30km for afrotype) and initial surface heat-flow values (Q_S) (taken from Buck, 1991, fig. 1). The large mantle plume under the right-hand afrotype taphrogen is from White and McKenzie (1989). The width of the magmatic arc in the upper cross section is about 200km; there is no vertical exaggeration in either section.

words, intracontinental taphrogeny leads to *koilogeny* (Spizaharsky and Borovikov, 1966, p. 113ff: from the Greek κοῖλος = hollow) (Fig. 2.4).

As we need to recognize in the past not only individual extensional structures, but also *patterns of structures* (i.e., whole taphrogens), in the following section, I present a hierarchical classification of rifts in the framework of taphrogens, which goes from pure geometry to dynamics, to enable us in our study of old rifts to be able to use data not only from individual rift fragments preserved in the record, but also from parts of larger patterns of rifts in relation to one another and/or to other structures such as koilogens, thalassogens and orogens.

Classification of Rifts

The classification of rifts offered here is concerned more with *groups of rifts* (i.e., taphrogens), than with individual rifts. It has three different categories that do not completely overlap, namely *geometric, kinematic* and *dynamic*. In the following, the three different categories are identified with their initials (i.e., *g, k,* and *d,* respectively). However, it is helpful to review, in passing, the three main kinds of individual rift types proposed by Buck (1991) that form in response to changes in crustal thickness, heat flow, and extension rate, to show how they relate to the classification here proposed.


NCh
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WG
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BT
SH
BR
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BR
CP
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AT
RG
CL
SL
KP
EASTERN NORTH AMERICAN RIFTS
TH
WH
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Cs
OS
AP
N
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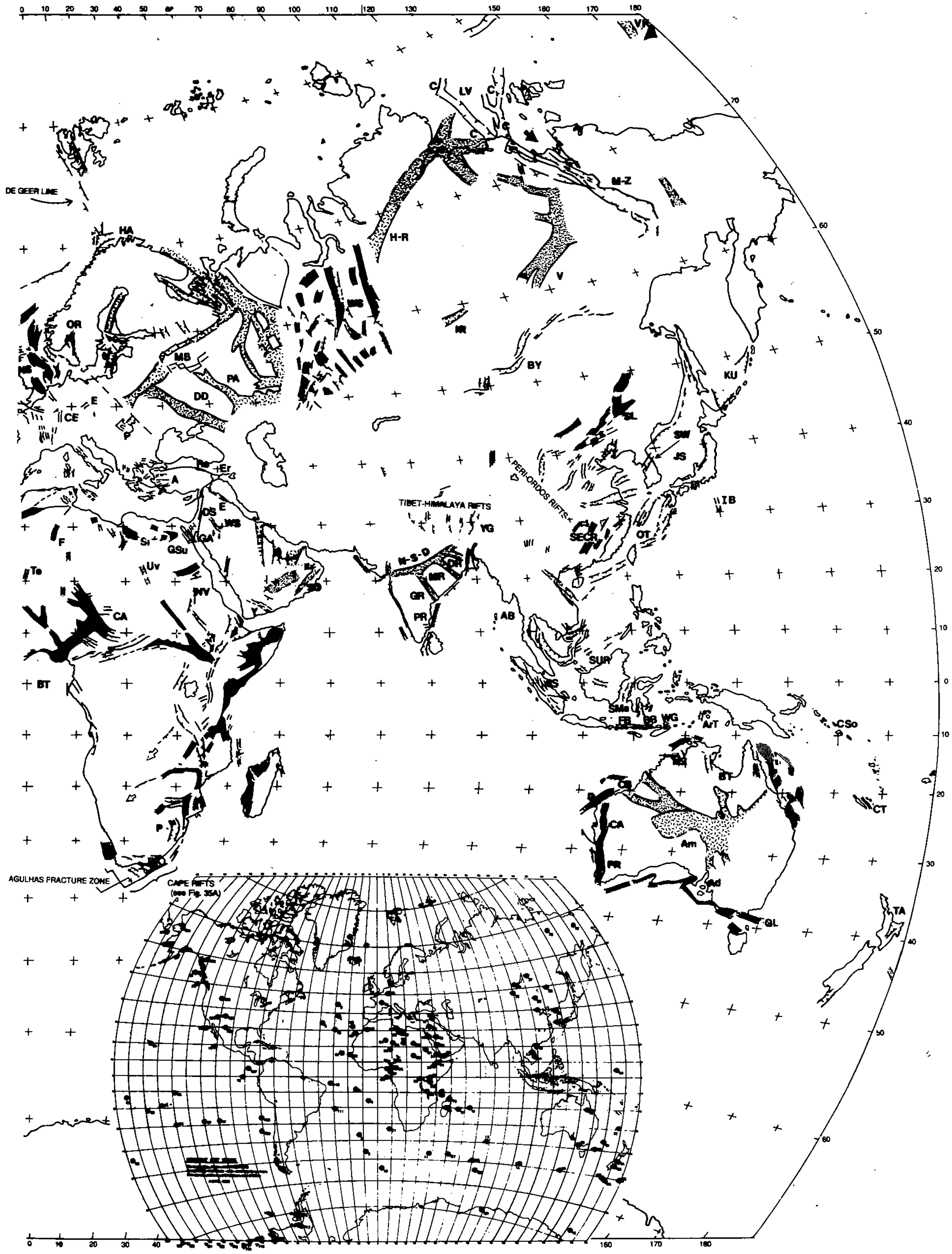
CE
G
LV
M
T-A
SL
B
AZ
P
AL
Ch
ESPIRITO SANTO
MESOZOIC SOUTH AMERICAN TAPHROGEN
SOUTH ARGENTINIAN RIFTS (see Fig. 13)
MALVINAS PLATEAU
CAINOZOIC
MESOZOIC
PALAEOZOIC
PRECAMBRIAN
PROPAGATING RIFT
PROPAGATING SPREADING CENTRE


DE GEER LINE
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TIBET-HIMALAYA RIFTS
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MR
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SUR
A
Er
DS
GSu
Uv
INV
CA
BT
Te
F
P
AGULHAS FRACTURE ZONE
CAPE RIFTS
(see Fig. 35A)
SMa
FB
BB
WG
ArT
CSo
CT
Am
Ad
GL
TA

Figure 2.2 World distribution of rifts, showing only those rifts that did not undergo enough later compression and/or strike-slip so as to obliterate their rift signatures. Owing to our heterogeneous knowledge of rifts of the world and to the scale, this map is more representative than exhaustive, and is intended to show that rifts indeed occur in all tectonic settings and that this has been so since the Archean! The map was updated using numerous sources, from a similar one in Burke et al. (1978). A few notable features of the map are the following:

1. Characteristics of taphrogens:

1.1. The map shows that taphrogens come in all scales ranging from singular, insignificant rifts, such as the Tesoffi rift in North Africa (Te) to immense rift complexes, such as the huge Mesozoic taphrogen of Argentina and Chile.

1.2. Taphrogens share plan-view characteristics with orogens. They exhibit curves in trend called *deflections* (e.g., pronounced deflection displayed by the Mid-Continent rift in Proterozoic basement of the United States). They can splay into horsetail-shaped *virgations,* as seen in the case of the East European rifts under the Moscow basin (MB); the virgations may be *one-sided,* as in the case of the Moscow virgation or they may be *two-sided,* as in the case of the smaller Central African rift system (CA). In East Africa, virgations are locally termed *divergences* (e.g., Daly et al., 1989). *Linkages* between two rift systems also occur, as the Kapuskasing rift (KP) meeting the Mid-Continent rift system under the Great Lakes area. Both linkages and virgations give rise to rift stars. *Taphroclines* are sharp bends in rift chains induced by bending after the formation of the rift chain, analogous to oroclines in orogenic belts. The best documented taphrocline is possibly formed by rifts that led to the opening of the Japan Sea, that became bent by nearly 90° as a result of the trap-door opening mechanism of the marginal basin. This involved an along-strike elongation as well, that thus makes the taphrocline really a *taphroclinotath*!

Unlike orogens, taphrogens may acquire roughly *equant* shapes, as exhibited by the West Siberian Rift System (WS) or the North Sea Mesozoic taphrogen (NS), or the Aegean rifts (A). Equant taphrogens also come in various sizes, as these examples show, and they usually result from rift clusters.

1.3. Rift clusters do not commonly seem to lead to ocean generation (i.e., to thalassogeny) because nearly all examples exhibited on this map have remained intact (e.g., West Siberian taphrogen (WS), Basin and Range taphrogen (BR), North Sea taphrogen (NS) or at least the major part of it, and Aegean taphrogen (A). This is true also for some of the better known rift clusters that have later fallen prey to orogeny (e.g., the early Paleozoic rift cluster in southern Tasman orogen in eastern Australia: B.C. Burchfiel, personal communication, 1977; Burke et al., 1978; Veevers, 1984). The immense Mesozoic taphrogen of South America is, in this regard, an exception, but here we may be witnessing the superposition of a rift chain of Late Jurassic - Early Cretaceous age on an older Mesozoic rift cluster!

1.4. More so than in orogens, in taphrogens, spatially and temporally persistent trends are easily discernible. In the Mesozoic Central African rift system, for example, northwesterly trends dominate. A similar trend dominates the large South American taphrogen. In both cases, these trends are a reflection of the grain of the basement fabric. Because rocks are weaker under tension than under compression, basement trends control taphrogeny more easily than they can orogeny.

1.5. Cenozoic collision-related rifts are almost invariably narrower than intra-plate or constructive-margin rifts.

1.6. If this map is compared with one showing age of crust, it will be seen that the younger a piece of crust, the narrower the rifts that form in it. I interpret this in terms of lithospheric thickness. Younger crust indicates thinner lithosphere and the thinner a piece of lithosphere is, the narrower are the rifts that disrupt it.

2. Characteristics of ocean margins related to taphrogens:

2.1. Most ocean margins of Atlantic type today appear to have grown out of Mesozoic taphrogens. In some cases, the parental taphrogen was a rift chain split down its axis (e.g., eastern African margin, central Atlantic margins of North America and Africa). In others, a rift cluster became severed along a line oblique to its dominant rift grain (e.g., in South Africa and in the Falkland spur of South America). This latter is either a result of shear separation (e.g., South Africa/Falkland Spur separation along the Agulhas fracture zone) or the superposition of a rift chain on an older rift cluster (as in the case of Argentina). We may thus distinguish two sub-groups within Atlantic-type continental margins, namely a *Central Atlanic type* (rift chain split down its axis) and a *South Atlantic type* (thalassogeny obliquely cuts older taphrogenic grain).

2.2. A consideration of the geometry of the major taphrogens (e.g., in Eastern Europe or in East Africa) shows how microcontinents might have been produced if all depicted rifts had formed oceans. Rift chains or rift nets commonly lead to roughly equant or at best "diamond-shaped" microcontinents. By contrast, rift clusters create, at the thalassogenic stage, more ribbon continents than anything else (e.g., Baja California).

2.3. Present continental margins of Atlantic type are overwhelmingly of Mesozoic age. Almost no Atlantic-type margin of Paleozoic or older age survives.

Key to abbreviations not mentioned above:

ASIA: AB = Andaman Basin; ArT = Aru trough; BB = Bone basin; BY = Baikal rift; CS = Central Sumatra basin; DR = Damodar rift; Er = Erzincan pull-apart rift; FB = Flores back-arc rift; GR = Godavari rift; H-R = Hantaj-Rbyninsk rift; I-B = Izu-Bonin rifts; Ir = Irkineev rift; KU = Kurile basin; MR = Mahanadi Basin; MW = Gulf of Mannowar rift; M-Z = Moma-Zoryansk rifts; N-S-D = Narmada-Son-Damodar rift lineament; OT = Okinawa Trough; PR = Pondicherry rift; Re = Reşadiye pull-apart rift; SECR = Southeast China rifts; SL = Song Liao rifts; SMa = South Makassar rift; SUR = Sundaland rifts; V = Vilyuy rift; WG = Wetar "graben" rift; Y-G = Yadong-Gulu rift.

EUROPE: CE = Central European "pack-ice-type" rifts and impactogens; DD = Dnyepr-Donetz aulacogen; E = Eger rift; HA = Hammersfest basin; OR = Oslo rift; P = Pannonian basin rifts; PA = Pachelma aulacogen.

AFRICA: BT = Benue Trough; Cs = Casamance rift; H = Haha rift; NV = Nile Valley faults; P = Pongola basin (world's oldest preserved rift: see Burke et al., 1985); Te = Tessofi rift.

ARABIA: (The gently compressed Syrian rifts are not shown) DS = Dead Sea pull-apart rift; E = Euphrates "graben" rift; GA = Gulf of Aqaba rift; GSu = Gulf of Suez rift; RA = Rub-al-Khali rift; SO = South Oman rift, WS= Wadi Sirhan rift, Y = South Yemen rift.

AUSTRALIA, NEW ZEALAND, OCEANIA: Ad = Adelaide aulacogen; Am = Amadeus basin; BT = Batten trough rift; CA = Carnarvon rift; CB = Canning basin rift (with the Fitzroy rift); CSo = Central Solomons trough (Russel Basin); CT = Coriolis trough; DR = Dampier rift; GL = Gippsland basin rift; PR = Perth rift; TA = Taupa rift.

NORTH AMERICA: AL = Alberta; AT = Athapuscow rift; B = Bathurst rift; BT = Belt rift; CL = Cambrian Lake; CP = Coppermine; GBB = Grand Banks basins; GG = Guatemalan rifts; KW = Keweenawan rifts (part of the Mid-Continent rift system); M = Marathon rift; MSS = Mississippi embayment rift (including the Reelfoot rift); RiG = Rio Grande rift; RG= Richmond Gulf rift; SL = Seal Lake; SN = Snake River; UN = Uinta rift; W = Wichita rift.

SOUTH AMERICA: AL = Altiplano-Puna rifts; AZ = Amazon rift; B = Barreirinhas rift; Ch = Chiquitos rift; M = Maracaibo rifts; P = Cordillera Blanca rift; SL = San Luis; T-A = Takutu-Apoteri rift.

Inset: A liberal interpretation of the active hot spots of the world (defined as "non-plate margin-magmatism": see Burke et al., 1978, Fig. 6.2.14), after the unpublished work of W.S.F. Kidd, S. Anderson, K. Burke, C. Chatelain, and C. White (Albany Global Tectonics Group, 1980). Other and somewhat simplified versions of this map were published in Burke et al. (1978, Fig. 6.2.4) and Turcotte and Schubert (1982, fig. 1-49). For a more conservative view of hot-spot distribution, see Crough and Jurdy (1980). Another map published by Vogt (1981) is only a slightly modified version of the Kidd et al. map. I follow here the view of Kidd et al., with the reservations expressed below.

Hot spots on or near divergent plate boundaries:

No.	*Name*	*Coordinates*	*Plates*
1	Jan Mayen	71N, 9W	EUR/NAM
2	Iceland	65N, 16W	EUR/NAM
3	45°	44N, 28W	EUR/NAM
4	Azores	38N, 27W	EUR/AFR/NAM
5	Colorado Seamount	34N, 38W	AFR/NAM
6	Ascension	8S, 14W	SAM/AFR
7	St. Helena	16S, 6W	AFR/SAM
8	Tristan/Gough	38S, 11W	AFR/SAM
9	Discovery Seamounts	46S, 8W	AFR/SAM
10	Bouvet	54S, 3E	ANT/AFR/SAM
11	Prince Edward/Marion	47S, 38E	ANT/AFR
12	Rodriguez	20S, 62E	AFR/IND
13	St. Paul/New Amsterdam	39S, 78E	ANT/IND
14	Naturaliste **Q**	48S, 105E	ANT/IND
15	Kangaroo **Q**	49S, 135E	ANT/IND
16	Balleny	67S, 163E	ANT/PAC/IND
17	Easter	27S, 109W	NAZ/PAC
18	Galapagos	0, 91W	NAZ/COC
19	Islas Revillagigedo	19N, 111W	PAC/NAM/COC
20	Explorer Seamount	49N, 132W	PAC/GOR
21	Bowie Seamount	53N, 136W	PAC/NAM/GOR
22	Harrat Er-Raha **Q**	27N, 36E	ARA/AFR
23	Aden	14N, 46E	ARA/AFR
24	Danakil	15N, 41E	AFR/ARA
25	Awash	10N, 41E	AFR/ARA

Hot spots on major active taphrogens:

26	S. Wonji	7N, 38E	AFR
27	Nakuru	0, 36E	AFR
28	Kilimanjaro	3S, 36E	AFR
29	Nyirangongo	1S, 29E	AFR
30	Rungwe	9S, 34E	AFR
31	NE		
	Ordos Plateau		
32	**Q**	40N, 114E	CHI
	Balagan Tas		
33	Indigirsky **Q**	67N, 144E	CHI
	Baikal/		
	Irkutsk **Q**	52N, 104E	EUR/CHI

Hot spots within plates:

34	Massif Central	46N, 3E	EUR
35	Vogelsberg **Q**	50.5N, 7.5 E	EUR
36	Ara Hangay		
37	**Q**	48N, 100E	CHI

38	Vitim Plateau	54N, 113E	CHI
39	Nen-Chiang	49N, 125E	CHI
40	**Q**		
	SW Hsing-An	45N, 115E	
41	**Q**		CHI
42	Soeul	39N, 127E	
43	Wonsan **Q**	33N, 126E	CHI
44	Cheju **Q**	21N, 110E	CHI
45	Hainan **Q**	16N, 107E	CHI
46	SE Laos **Q**	13N, 108E	CHI
	SE Vietnam **Q**	1N, 110E	CHI
47	W. Borneo **Q**		CHI
	Central	1N, 110E	
48	Borneo **Q**		CHI
49	Vema	33S, 8E	
50	Seamount	33N, 17W	AFR
51	Madeira	28N, 15W	AFR
52	Canary	31N, 7W	AFR
53	Anti-Atlas	16N, 24W	AFR
54	Cape Verde	15N, 18W	AFR
55	Dakar	18N, 9E	AFR
56	Air	24N, 6E	AFR
57	Ahaggar	32N, 13E	AFR
58	Tripoli	29N, 15E	AFR
59	Jebel Sawda	28N, 17E	AFR
60	Haroudj	23N, 11E	AFR
61	Ih Ezzane	23N, 18E	AFR
62	Tibesti	23N, 25E	AFR
63	Jebel Uweinat	18N, 33E	AFR
64	Bayuda	14N, 25E	AFR
65	Jebel Marra	11N, 12E	AFR
66	Biu	10N, 9E	AFR
67	Jos Plateau	6N, 10E	AFR
68	Adamawa	6N, 13E	AFR
	Ngaoundere	4N, 9E	AFR
69	Cameroon		AFR
70	Sao Tome/	0, 6E	
71	Annobon	12S, 44E	AFR
	Comores	14E, 49E	AFR
72			AFR
		20S, 47E	
73			AFR
		21S, 56E	
74			AFR
75		31N, 29W	
76		22N, 42E	AFR
77		26N, 41E	ARA
78		33N, 37E	ARA
			ARA
79		37N, 39E	ARA
		20S, 145E	
80	Cap dÕAmbre	38S, 143E	IND
	Central Madagascar		IND
81	Reunion/	39S, 155E	
	Mauritius		IND
82	Great Meteor Seamount	36S, 160E	
83	Mecca **Q**	29S, 168E	IND
84	Medina **Q**	46S, 173E	IND
	Damascus **Q**		PAC
85	Karacada¤ **Q**	44S, 174E	
86	W. **Q**ueensland	44S, 176W	PAC
	W. Victoria		PAC
87	Tasmantid Seamounts	52S, 169E	

	Lord Howe Rise		PAC
88	Norfolk	50S, 166E	
	E. Otago **T**		PAC
89	Banks Peninsula **T**	50S, 179E	
90	Chatham Id. **T**	14S, 172 W	PAC
91	Campbell Id. **T**	5N, 164E	PAC
92	Auckland Id. **T**	20N, 155W	PAC
93	Antipodes Id. **T**	29N, 118W	PAC
94	Samoa	22S, 158W	PAC
95	Caroline	18S, 148W	PAC
96	Hawaii	11S, 138W	PAC
	Guadalupe		PAC
97	Rarotonga	29S, 140W	
	Tahiti		PAC
98	Marquesas	27S, 120W	
99	Macdonald Seamount	67N, 165E	PAC
100	Pitcairn/	64N, 170W	NAM
101	Gambier	61N, 165W	NAM
	Anjuisky		NAM
102	St. Lawrence	66N, 160W	
103	Nunivak	58N, 131W	NAM
	Seward Peninsula		NAM
104	Mt. Edziza **T**	44N, 112W	
105	Yellowstone/Snake River **S**	35N, 112W	NAM
106	Flagstaff	36N, 106W	NAM
	Santa Fe **S**		NAM
	Rio Grande/Big Bend **Q**, **S**		
107	Cocos	30N, 105W	
108	Isla San Felix	5N, 87W	NAM
109	Isla Juan Fernandez	26S, 80W	COC
	Fernando de Noronha		NAZ
110	Trinidade/	34S, 81W	
	Martin Vaz		NAZ
111	Peter 1st Id.	4S, 32W	
	Merick Mts.		SAM
112	Alexander	21S, 28W	
113	Hudson/	69S, 91W	SAM
114	Jones Mts.	75S, 72W	ANT
115	Murphy/	72S, 70W	ANT
	Crary Mts		ANT
116	Mt. Siple	74S, 97W	
	Ames/Flood Ranges		ANT
117	Executive Committee Range	77S, 115W	
118	Fosdick Mts.	73S, 126W	ANT
	Mt. Early		ANT
119	Mt. Erebus	76S, 135W	
	Mt. Melbourne/		ANT
	Adare Pen.		
120	Gaussberg	78S, 126W	
121	Heard	77S, 144W	ANT
122	Kerguelen	87S, 153W	ANT
123	Crozet	78S, 167E	ANT
			ANT
124		73S, 170E	
125		67S, 89E	ANT
126		53S, 74E	ANT
127		49S, 70E	ANT
		47S, 52E	ANT
			ANT

Notes after names

Q: Questionable existence of hot spot; **S**: Associated with small taphrogen; **T**: Secondary extension in wide zone associated with major transform fault.
Plate abbreviations: EUR: Eurasia; CHI: China; AFR: Africa; ARA: Arabia; IND: India; PAC: Pacific; NAM: North America; COC: Cocos; NAZ: Nazca; SAM: South America; ANT: Antarctica.

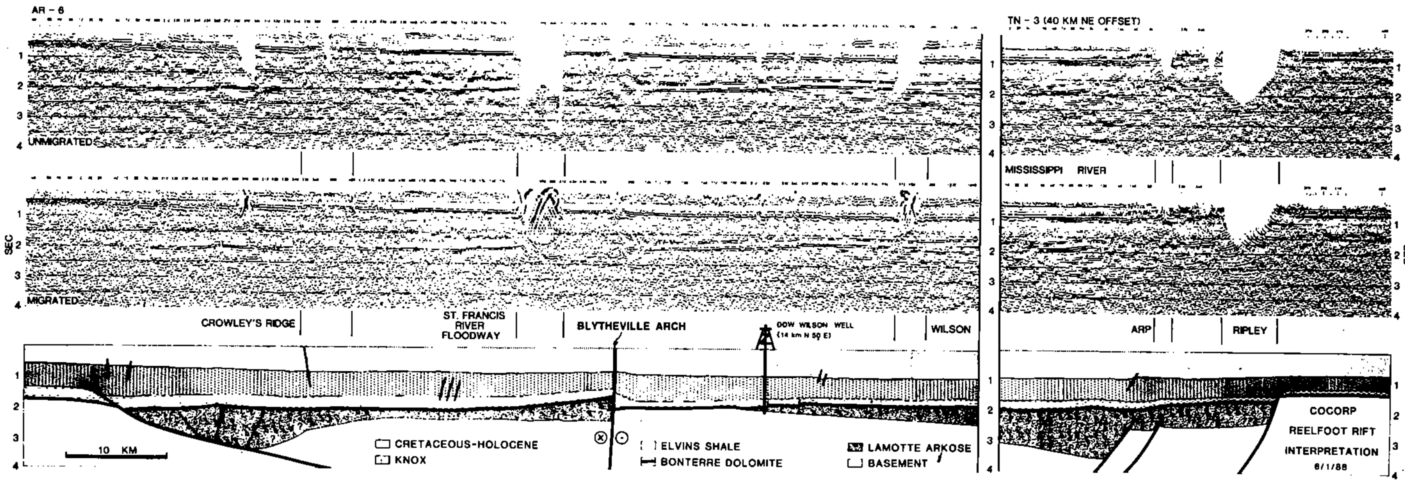


Figure 2.3A Deep seismic-reflection profile with interpretation across late Precambrian(?)/early Paleozoic Reelfoot rift buried beneath Cretaceous to Holocene strata of northern Mississippi embayment in southern United States. Rift was structurally and magmatically reactivated in Late Paleozoic and Late Cretaceous, and remains seismically active today, as shown by destructive 1811-1812 New Madrid earthquakes and contemporary seismicity. (Reproduced with permisson from Nelson and Zhang, 1991, Fig. 4.)

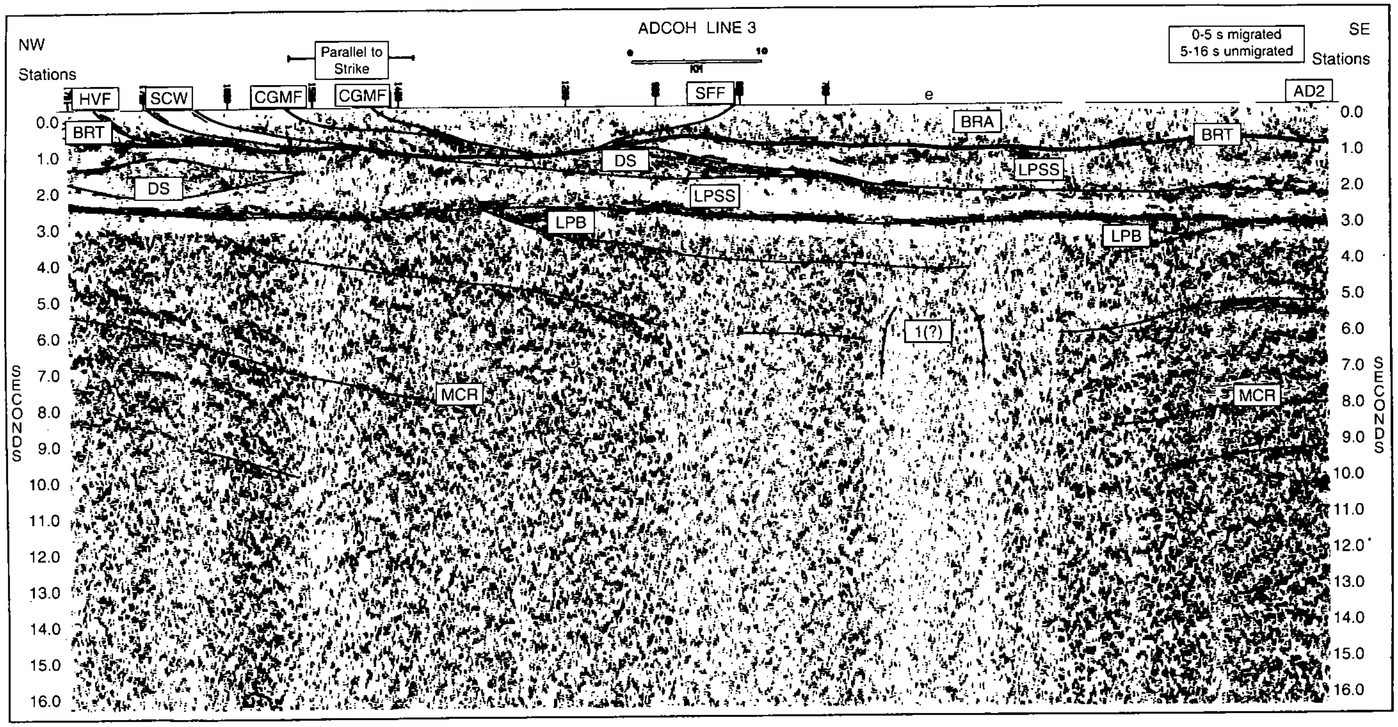


Figure 2.3B Deep seismic-reflection profile with interpretation across southern Appalachians along Appalachian Ultra-Deep Core Hole (ADCOH), line 3 showing Late Proterozoic/Early Paleozoic rifts (labeled as "LPB", i.e., late Proterozoic basin) now tectonically buried beneath Appalachian allochthons and parautochthon. Abbreviations are BRT: Blue Ridge thrust; HVT: Hayesvile fault; CGMT: Chunky Gal Mountain fault; SFF: Shope Fork fault; SCW: Shooting Creek window; BRA: Blue Ridge allochthon; LPSS: lower Paleozoic shelf strata (parautochthon); I(?): intrusion; DS: duplex structure; MCR: mid-crustal reflections. (Reproduced with permisson from Hubbard et al., 1991, Fig. 4.)

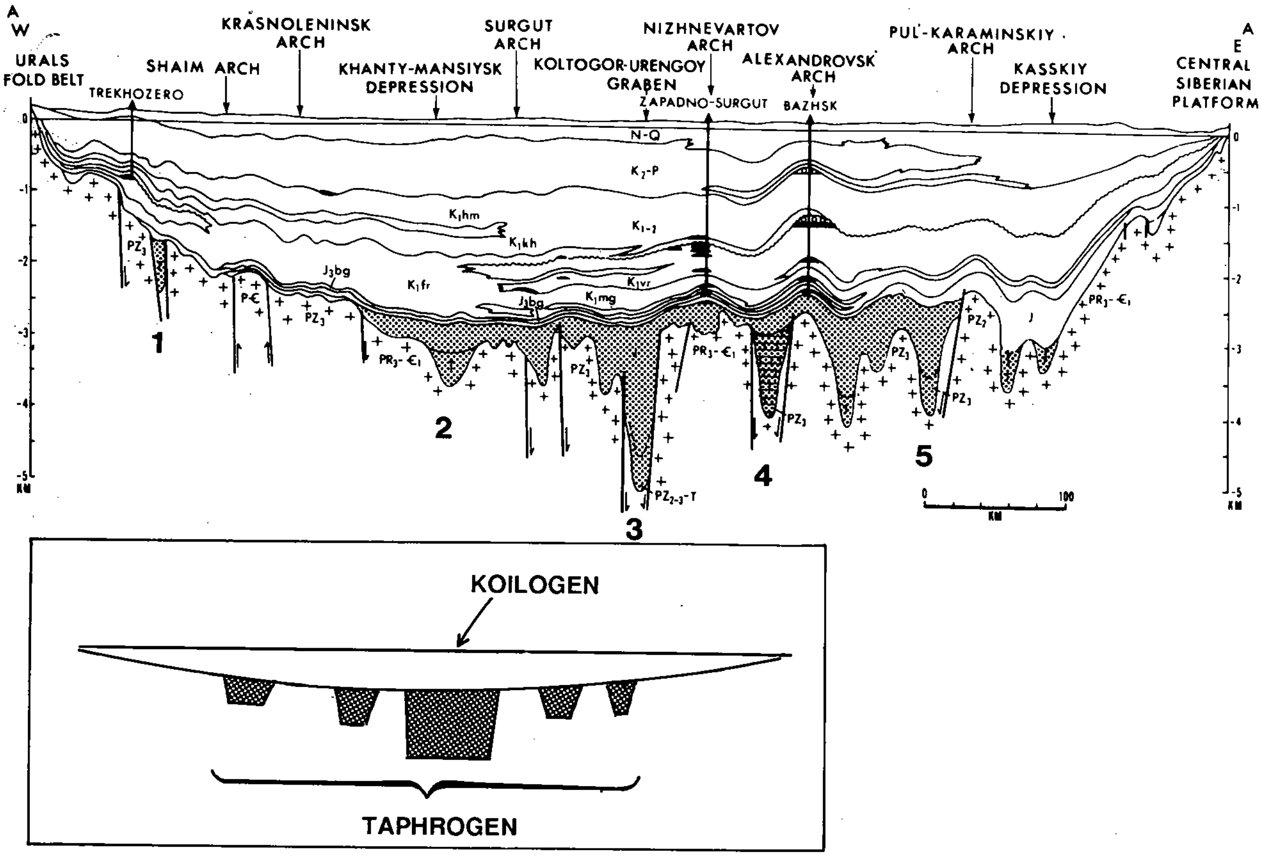


Figure 2.4 E-W schematic cross section across southern part of West Siberian basin, showing middle to late Mesozoic-Cenozoic *koilogen* nested on a late Paleozoic - early Mesozoic *taphrogen*, consisting of several rifts (major ones labeled 1 through 5). Sedimentary rocks filling rift basins (i.e., those deposited during taphrogen stage) are stippled. Location of the section (A-A') is shown in Fig. 2.8B. PC: Precambrian; PR_3: Vendian (Neoproterozoic); PZ_2: middle Paleozoic; PZ_3: upper Paleozoic; C_1: lower Carboniferous; T: Triassic (may include some Permian); J: Jurassic; J_3bg: Upper Jurassic Bazhenov and Georgiev formations, forming important source rocks; Kfr: Fedorov Formation; KKl: Kulomzin Formation; Kmg: Megion Formation; Kvr: Nizhnevartov Formation; Kkh: Kharosoimsk Formation; Khm: Khantimansisnsk Formation; K_1: Lower Cretaceous; K_2: Upper Cretaceous; P: Paleogene; N: Neogene; Q: Quaternary. (Modified with permissọn from Meyerhoff, 1982, Fig. 23.) Inset shows the "bovine head" model for the formation of koilogens through taphrogeny.

Structural Classification of Rifts (Fig. 2.1)

When buoyancy forces resulting from crustal-thickness differences created by extension can overcome tensional stresses, rifting loci migrate about a taphrogen; if not, then rifting occurs along a single locus. When, in addition, lower crust can flow faster than extension can augment crustal-thickness differences (see esp. Block and Royden, 1990, and Wernicke, 1991, for lower-crustal flow under extension), extension remains localized in the upper (i.e., non-flowing) crust, but is diffused in the lower (i.e., in the readily flowing) crust. For a discussion on the possible precise location and nature of these parts, see Wernicke (1991). These differences lead to the following three modes of rifting (Buck, 1991):

1) Briefly, when heat flow is high (about 100 mW/m²) and the crust thick (about 60 km), fairly rapid extension (about 1 to 2 cm/y) leads to localized upper-crustal extension, compensated at depth by crustal flow distributing thinning in an area larger than the surficial extension. Buck (1991) calls this "core-complex mode." A taphrogen extending in a core-complex mode is herein termed an *amerotype taphrogen* from its type locality

in the western United States (Fig. 2.1: Coney, 1979, 1980a; Armstrong, 1982). Amerotype rifts are poor sediment receptacles, for they do not create large depocenters, associated localized extension being confined to the upper crust, and the local changes in crustal thickness being small (leading to small isostatic subsidence). For the same reason, such taphrogens must lead to very wide, but shallow koilogens.

2) When heat flow is lower (about 80 mW/m^2) and the crust thinner (about 45 km or so), rapid to very rapid stretching (1 to almost 10 cm/y) leads to rifting in a wide area (a few hundreds of km. wide). Buck (1991) calls this "wide-rift mode" extension, which leads to what is here called *aegeotype* taphrogens (Fig. 2.1), after its best example in the Aegean Sea and its surroundings. Aegeotype taphrogens create rifts deeper and wider than the amerotype ones, but still not nearly as deep as the afrotype ones (see below). By contrast, the aegeotype taphrogens are followed by large and deep koilogens, such as the West Siberian Lowlands, the "younger internal basins" of eastern Australia, the North Sea, and the Song Liao basin in northeastern China (Fig. 2.2).

3) Normal-thickness crust (about 30 to 35 km) and heat flow (about 60 mW/m^2) lead to the generation of a narrow locus of rifting, when low extension rates (generally less than 1 cm/y) are applied to the lithosphere (Buck, 1991). I call this type of rifting *afrotype*, after its most spectacular representative in East Africa (Fig. 2.1). Afrotype taphrogens create deep and wide rifts, but small koilogens afterwards.

Geometric Classification of Rifts (Fig. 2.5)

Rifts display five kinds of patterns in map view (Şengör, 1983). From simplest to more complex, these are:

g1) Solitary rifts

Solitary rifts form small, fairly insignificant and rare taphrogens and are extremely difficult to ascertain in the geological record because it is often impossible to tell whether a given rift fragment is isolated or part of a larger area of rifting. The Cambrian Tesoffi Rift (Fig. 2.2; Kampunzu and Popoff, 1991) may be one such solitary rift.

g2) Rift stars

Rift stars (Fig. 2.6) form when more than two rifts radiate away from a common center, building a fairly equant taphrogen. Rift stars are common features of the structural repertoire of our planet today (e.g., the numerous rift stars called triple junctions by Burke and Whiteman, 1973, and Burke and Dewey, 1973; also, see Burke et al., 1972) and they seem to have been so in the past as well (see Burke and Dewey, 1973; Burke, 1976, 1977a; Şengör, 1987a).

g3) Rift chains

When several rifts are aligned end-to-end along linear/arcuate belts of rifting, they form *rift chains*. The East African Rift System constitutes the best known active rift chain in the world (Fig. 2.7A). A similar rift system appears to have been responsible for the opening of the Atlantic Ocean, which constitutes the best-studied fossil example of a rift chain (Fig. 2.7B: Burke, 1976). In general, major rift chains do not fail in toto to form ocean, as the example of the rift chain associated with the Atlantic Ocean shows. An ancient rift chain that did not lead to ocean opening, on the other hand, is represented by the Keweenawan (or the "Midcontinent") rift located mainly in the basement of the U.S. Interior Lowlands (Fig. 2.2), but its length is considerably shorter than either the Atlantic rift chain or the extant East African one (Van Schmus and Hinze, 1985; Gordon and Hempton, 1986).

Solitary rifts, rift stars and rift chains commonly form from afrotype rifts, although aegeotypes are also known to house rift stars (e.g., the rifts possibly generated by migration of the Yellowstone hot spot in the Columbia basalt province plus the Snake River plain: see Davis, 1977, esp. Fig. 2R C–6). A fossil example may be the Jurassic-Cretaceous aulacogen-generating rifts superposed on an earlier Triassic-Jurassic rift cluster in southern South America (Fig. 2.2; for some possible west Siberian cases, see Milanovsky, 1987, esp. Fig. 12).

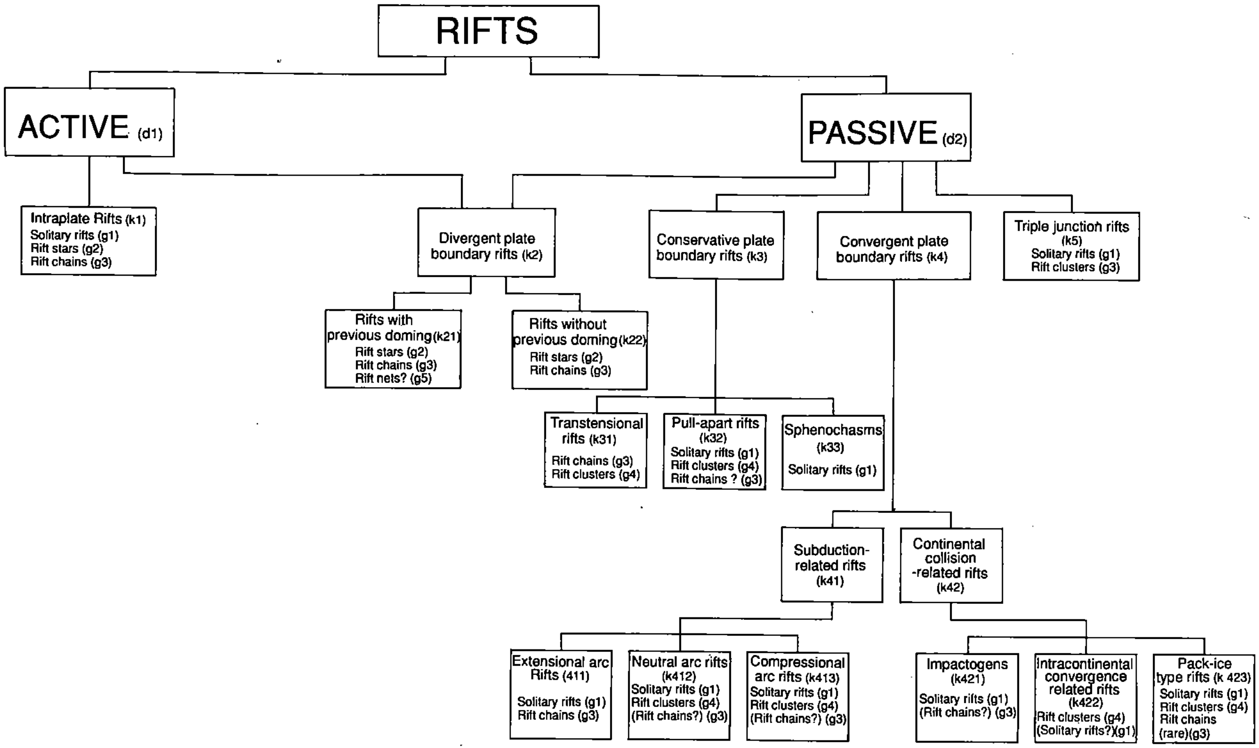


Figure 2.5 Classification of rifts

g4) Rift clusters

When several sub-parallel rifts occur in roughly equant areas, they are said to form a *rift cluster* (Şengör and Burke, 1978, p. 419). The two best-known active rift clusters in the world are the Basin and Range extensional area in the Great Basin (Fig. 2.8A: Gilbert, 1928; Becker, 1934; King, 1977; Stewart, 1978) and the Aegean Sea and surrounding regions ("Aegea" of Le Pichon and Angelier, 1981; McKenzie, 1978b; Şengör, 1978, 1982; Dewey and Şengör, 1979; Şengör et al., 1985; Şengör, 1987c). An equally impressive, but less well-known active rift cluster is the one in Tibet (Fig. 2.2; Tapponnier et al., 1986). The rift cluster underlying the West Siberian Lowlands is the most impressive fossil example (Fig. 2.8B: Rudkevich, 1976; Surkov and Djero, 1981; Meyerhoff, 1982). The rift clusters commonly form from aegeotype rifts, but some amerotype rifts may also lead to rift-cluster formation (e.g., the early Basin and Range).

g5) Rift nets

Rift nets constitute a rare pattern, which comes about when rifts form a roughly checkered pattern, as in the Neoproterozoic basement of the East European platform (Fig. 2.2) or in the late Mesozoic in central North Africa (Fig. 2.2). They resemble chocolate-bar boudinage and may have a similar origin, but more commonly rift nets form in complex and rapidly shifting stress environments in which dominant extension directions change rapidly. Many rift nets, in fact, may represent two superposed rift clusters. All three types of rifts may generate rift nets, but genuine rift nets (i.e., those not formed by the crossing of two rift clusters) are generally formed from afrotype members.

Kinematic Classification of Rifts (Fig. 2.5)

As rifts occur during all stages of the Wilson Cycle, the kinematic characteristics of the plate boundaries may be taken as bases for classifying them according

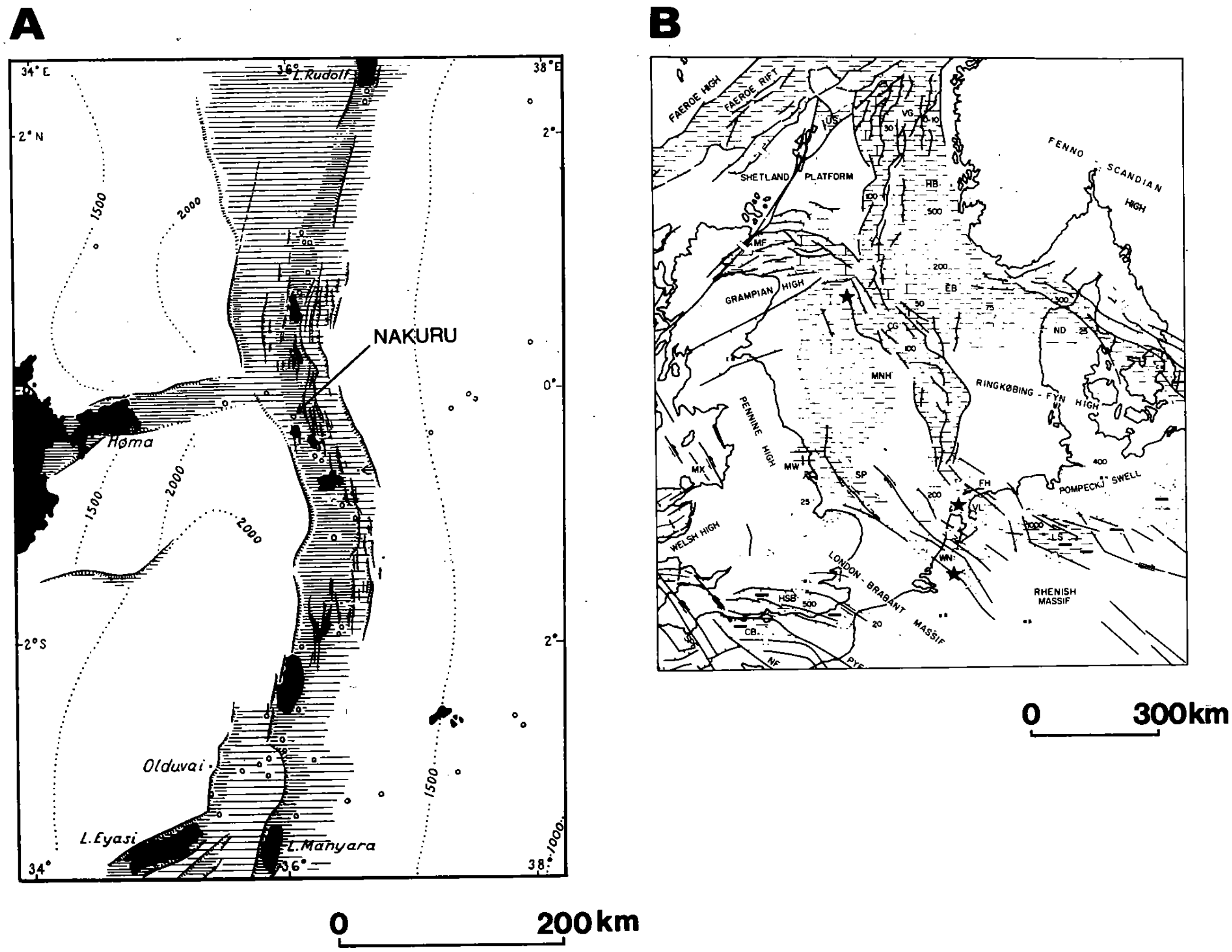


Figure 2.6 Two rift stars that have formed through different mechanisms:

A. Nakuru rrr junction (see also Burke and Whiteman, 1973, p. 743 and Fig. 5), at which two active arms forming Gregory rift, and Kavirondo rift, which ceased activity in Pliocene, meet. This junction formed through doming-rifting mechanism (path d1-k2-k21-g3). (Modified with permission from Shackleton, 1955, Fig. 1.)

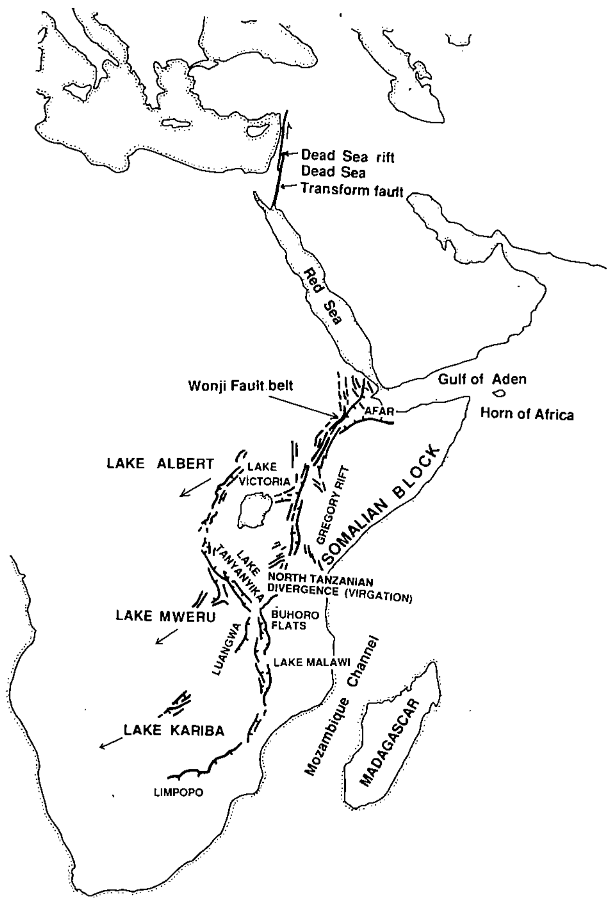
B. Berriasian-Valanginian (Early Cretaceous) paleotectonics of North Sea rift star, whose arms were formed from Viking Graben (VG), Central Graben (CG), and Moray Firth rift (MF). This rift star formed mainly in latest Permian - Early Triassic through superposition of a pull-apart system on a previously formed intracontinental-convergence-related rift system (path d2-k4-K42-k422-?g1 superposed by the path d2-k3-k32-g4). (Reproduced with permission from Ziegler, 1990, Fig. 1.13.)

to the environment of the overall displacement and strain field in which they form. There are three types of plate boundaries, plus the plate interiors, with which four types of rift families correspond. In addition, problems of compatibility arise around some unstable triple junctions, commonly owing to involvement of hard-to-subduct buoyant lithosphere. Some of these problems lead to complex rifting that should be treated separately from the other four classes, thus creating a fifth kinematic class, herein called triple-junction rifts.

k1) Intraplate rifts (exclusively afrotype)

Rifts surrounded entirely by undeformed lithosphere occupy this category. Such rifts are usually solitary, small and rare, and are difficult to detect in the geological record. Some active examples are found in the northeastern United States in the Lake George and

A



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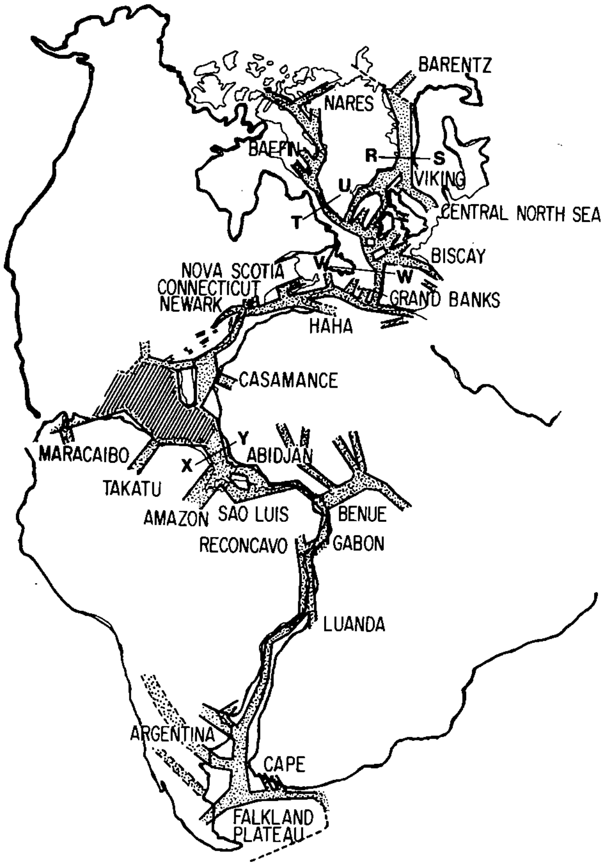


Figure 2.7 Rift chains:
A. Taphrogen of Great Rift Valley of East Africa as the best example of an active rift chain. Modified from Dixey (1956, Fig. 1) using more recent sources. Note how various branches of the rift chain are connected through rift stars. Compare this map with Fig. 2.11B. Also, compare it with one showing the distribution of sutures in a major collisional orogen (e.g., in the Tethysides: see , Şengör et al., 1988, Fig. 3) in order to have a better idea of the origin of various map views of former continental margins and microcontinents now caught up in an orogenic collage (see also Dewey and Burke, 1974; Scrutton, 1976). Arrows in central Africa indicate directions of rift-tip propagation after Girdler (1991, Fig. 1b).

B. Sketch map illustrating dominantly Mesozoic taphrogenic complex in western Pangea, forming a large, diachronous rift chain that eventually led to thalassogeny creating the present Atlantic Ocean. Many small rift basins and horsts could not be shown at this scale. Rifts developed between 210 and 170 Ma in area between lines V-W and X-Y. Oblique shading marks place from which continental material was removed prior to deposition of basal Jurassic salts on young ocean floor of Gulf of Mexico (see Burke et al., 1984, for a review of identity and placement of these continental fragments). Rifts south of line X-Y formed between about 145 and 125 Ma. Rifts north of line T-U formed about 80 Ma, and rifts north of R-S, 80-60 Ma. North of line R-S, Devonian, Permian, Triassic, and Jurassic taphrogenic episodes preceded Cretaceous taphrogeny that finally led to thalassogeny. (Modified from Burke, 1976, Fig.1.)

Lake Champlain rift structures (e.g., Burke, 1977). The Tesoffi rift in North Africa may be one fossil example (Fig. 2.2; Kampunzu and Popoff, 1991).

k2) Rifts associated with divergent plate boundaries (commonly afrotype, but may evolve from amero- or aegeotypes to afrotype just before the onset of thalassogeny)

These rifts form as a direct consequence of plate separation along nascent divergent boundaries. All rifts along the East African Rift system (Figs. 2.2 and 2.7A) belong to this category. In fact, along the Wonji fault belt in the Ethiopian rift segment, sea-floor spreading may have just commenced by axial dike injection (Mohr, 1971, Fig. 6). The latest Paleozoic-early Mesozoic grabens in eastern Canada and the

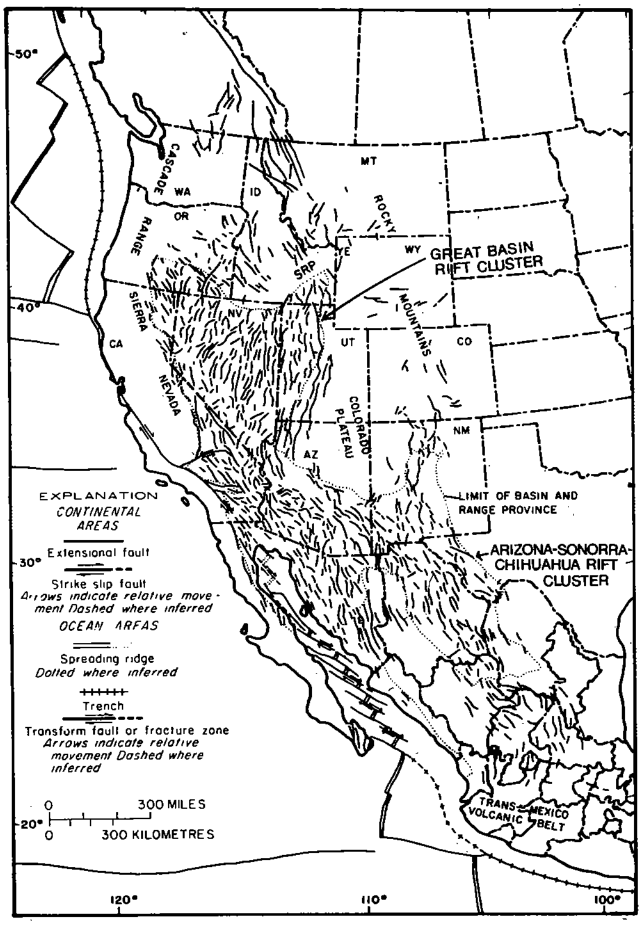
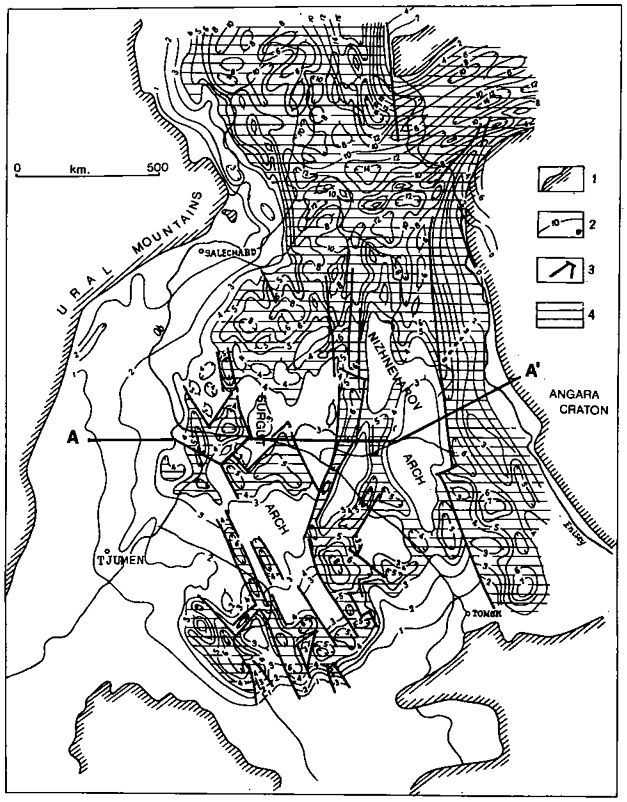


Figure 2.8 Rift clusters:
A. Distribution of late Cenozoic extensional faults and a few strike-slip faults in western North America defining Basin and Range province *sensu lato*. The whole area consists of one immense rift cluster, but it is possible to define two smaller clusters, namely the Great Basin cluster (Basin and Range Province *sensu stricto*: Gilbert, 1928; see also, King, 1977, Fig. 100) and the Arizona-Sonora-Chihuahua cluster. Key to abbreviations: WA, Washington; OR, Oregon; CA, California; CO, Colorado; ID, Idaho; MT, Montana; WY, Wyoming; NV, Nevada; UT, Utah; AZ, Arizona; NM, New Mexico; SRP, Snake River Plain; YE, Yellowstone. (Reproduced with permission from Stewart, 1978, Fig. 7.)



B. Sketch map of major structures of West Siberian basin at the top of the basement (i.e., pre-Triassic). Legend: 1: boundary of basin; 2: isobaths; 3: faults of the taphrogenic stage (i.e., latest Permian to medial Jurassic, forming a rift cluster); 4: basement areas formed during the middle Paleozoic through (?) Early Triassic (reproduced with permission from Aleinikov et al., 1980, Fig. 7). Note that West Siberian rift cluster is similar in size to Great Basin cluster. For cross section AA', see Fig. 2.4.

eastern United States constitute perhaps the best-known fossil examples of this category of rift structures (Fig. 2.7B) (Benson and Doyle, 1988; Hutchinson and Klitgord, 1988; Manspeizer, 1988; but see also Swanson, 1982). This category of rifts may be further subdivided into two classes as follows:

k21) Rifts that form following an episode of doming. The divergent boundary along which rifts form is, in this case, preceded by an episode of lithospheric doming. The East African Rift Valleys are outstanding extant examples of such a situation (cf., Thiessen et al., 1979). Rifts of Mesozoic age on the Atlantic margins of the Iberian peninsula yield evidence for a comparable situation (Wilson, 1975).

k22) Rifts that form with no pre-rift doming. The late Cenozoic Laptev-Moma-Zoryansk rift system probably represents a recent example (Fujita and Cook, 1990). A good fossil example is the rifting of the Alpine Neo-Tethys in the earlier Mesozoic.

k3) Rifts that form in association with conservative plate boundaries

Conservative boundaries are, by definition, those along which neither extension nor shortening takes place. However, various reasons conspire to induce both extension and shortening along considerable stretches of these boundaries (Christie-Blick and Biddle, 1985; Sylvester, 1988). Rifts along conservative plate boundaries form in three different settings:

k31) Transtensional conservative boundaries. If a conservative boundary is leaking because of a component of extension, it is called transtensional (Harland, 1971, p. 30 and esp. Fig. 2). Many active rifts have a transtensional component (e.g., the Gulf of Suez: Fairhead and Stuart, 1982, Fig. 1; the East African Rift System: Rosendahl, 1987, Fig. 1d) and fossil examples of such rifts may be recognized largely through the structures they contain (e.g., see Chorowicz et al., 1989, for an active example and Beauchamp, 1988, for a fossil one) or from their former bounding-transform-fault endings (e.g., see Tiercelin et al., 1988; Chorowicz, 1989; Daly et al., 1989; Chorowicz and Sorlien, 1992, for active examples; Şengör, 1990b, for a fossil example).

k32) Pull-apart basins along conservative boundaries. Major strike-slip faults, the main structural expression of conservative plate boundaries, commonly have bends along them that either facilitate ("releasing bends": Crowell, 1974b, Fig. 3) or obstruct ("restraining bends": Crowell, 1974b, Fig. 3) movement along the strike of the fault. These bends may be primary, related to the initial nucleation of the fault (e.g., Tchalenko, 1970; Merzer and Freund, 1974) or secondary, formed through structural modifications imposed on a preexisting fault and/or system of faults (e.g., Merzer and Freund, 1975; Mann et al., 1983, esp. Fig. 3; Şengör et al., 1985, esp. Figs. 10–12). In both cases, extensional basins form along the releasing bends, in which the magnitude of extension equals the magnitude of cumulative strike-slip offset along the strike-slip fault since the formation of the releasing bend. Such basins are called "pull-apart basins" after Burchfiel and Stewart's (1966) apposite appellation, but the concept is much older. These basins come in all forms and shapes, notwithstanding the claim by Aydin and Nur (1982) that they display a constant aspect ratio at all scales. Some, especially the small ones, display a thermal regime colder than other kinds of rifts owing to the proximity of cold bounding-fault walls (e.g., Pitman and Andrews, 1985) that may also contaminate their magmatic signature (e.g., Baş, 1979). One of the sedimentologically best-studied examples is the Dead Sea rift in the Middle East (Fig. 2.2; Manspeizer, 1985; for a recent alternative view, see: Ben-Avraham and Zoback, 1992; Ben-Avraham, in press; Ben-Avraham and Lyakhovsky, in press).

k33) Sphenochasms. Not all basins created by secondary extension associated with strike-slip faults are pull-apart basins. Some represent tears caused by either an asperity or differential drag along the strike-slip fault in one of the fault walls, in which the amount of extension changes from a maximum along the fault to zero at the pole of opening of the tear basin. Carey (1958, p. 193) called such wedge-shaped rifts that open towards a major strike-slip fault *sphenochasms* (from the Greek σφεν = corner , and χαω = to yawn). Highly seismic active examples of sphenochasm generation are seen in the Balkan Peninsula, between Albania and Thrace, as a consequence of the North Anatolian fault's attempt to tear Albania away from the rest of the peninsula (cf., Dewey and Şengör, 1979, esp. Fig. 1A). Another active example is constituted by the grabens around the Ordos block (Fig. 2.2; see Molnar and Tapponnier, 1975; Ma et al., 1982) and the Rio Grande rift in the western United States (Fig. 2.2; e.g., Eaton, 1979, esp. Fig. 7; for a recent overview see Ingersoll et al., 1990; also see Eardley, 1962, Fig. 31.22).

k4) Rifts that form in association with convergent plate boundaries

A large family of rifts forms in association with convergent plate boundaries. In this group, a first-order subdivision is between rifts associated with subduction zones and rifts associated with continental collision, although this may artificially split some genetic groups, such as those rifts that presumably form due to tension generated by excessive crustal thickening (e.g., Dewey, 1988; England and Houseman, 1988).[1] The usefulness of the present grouping is that it enables a rapid overview of the presently active rift environments and comparison with past ones.

k41) Rifts associated with subduction zones. Three separate environments of rifting associated with subduction zones correspond with three different types of arc behavior, namely, extensional, neutral, and compressional arcs (Dewey, 1980; Şengör, 1990a).[2] In these environments, an enormous variety of rifts forms and many evolve into oceans. In the following discussion, I consider only those that fail to generate oceans and get preserved as fossil rifts.

k411) Rifts associated with extensional arcs. Once an arc begins extending, it generally splits along the magmatic axis (if such an axis is already in existence) and forms a small rift chain (Karig, 1972, 1974; Klaus et al., 1992a) (see Chapters 7 and 8). Such a situation is today known from both the Okinawa rift (Fig. 2.2; Lee et al., 1980; Kimura, 1985; Letouzey and Kimura, 1985; Viallon et al., 1986; Kimura et al., 1988), and the Izu-Bonin arc system (Fig. 2.2; Klaus et al., 1992a), where marginal basins are in the process of rifting. Such rifts generally do not get preserved intact in the geological record, both because of the vicissitudes of the tectonic evolution of arcs involving common changes of behavior (see esp. Dewey, 1980; Şengör, 1990a), and because of later collisions with other arcs or continents. Preservation of rifts associated with extensional arcs in an uncompressed state takes place commonly when the associated arc switches from extensional behavior to neutral behavior.

Extensional arcs may also generate strike-slip faults parallel with the direction of extension, separating arc segments with different rates of extension. In such cases, some rifts in the less extended parts may be left behind and end up trending into fully developed marginal basins. The Neogene to Recent Seoul-Wansan graben in North Korea (Fig. 2.2) may be such a rift related to the opening of the Japan Sea marginal basin (Burke, 1977b).

k412) Rifts associated with neutral arcs. Neutral arcs are defined to have neither shortening nor extension across them. Therefore, the only rifts that form in neutral arcs are those associated with arc-parallel strike-slip faults and may be classified in exactly the same way as the rifts that form along conservative plate boundaries. More complex rift basins may form along such arc-parallel strike-slip faults, if the sliver plate in the forearc area (Jarrard, 1986b, p. 235; "forearc plate" of Woodcock, 1986) disintegrates and its various pieces rotate about vertical axes (e.g., Jesinkey et al., 1987; Beck, 1988).

Pull-apart basins in arcs are difficult to recognize. None of the major active strike-slip faults located in arcs has well-developed pull-apart basins along them (e.g., Median Tectonic Line in Japan, the Atacama fault in the Andes, or the Philippine fault in the Philippine Archipelago), except the Andaman Basin that connects the right-lateral Sagaing and the Semangko faults and that is likely floored by oceanic crust (cf., Hamilton, 1979). Fossil and relatively undeformed examples of such basins have been inferred and mapped, however (e.g., the Chuckanut, Puget, and Swauk basins in Washington, U.S.A.: Fig. 2.2; Johnson, 1985).

Sphenochasms along strike-slip faults in arcs are rarer still. Davis et al. (1978) have discussed two possible examples, the more recent of which may have created the "Columbia Embayment" (Fig. 2.2; Rogers and Novitsky-Evans, 1977a,b) by motion along the Straight Creek fault in the latest Cretaceous and the earliest Cenozoic.

K413) Rifts associated with compressional arcs. In compressional arcs, crust commonly thickens and lithosphere thins, both by heating and by eventual delamination (Isacks, 1988; Şengör, 1990a). The arc becomes shortened across, and elongated along, its trend (see esp. Dewey and Lamb, 1991). This elongation commonly generates rifts at high angles to the trend of the arc (e.g., in the Altiplano: Mercier, 1981; Mercier et al., 1991; in Crete: Le Pichon and Angelier, 1979).

In the Andes of Peru, however, Dalmayrac and Molnar (1981) have observed normal faulting parallel with the trend of the arc and ascribed this to gravitational stresses set up by the potential energy of the thickened crust. Later, Jordan et al. (1983a,b) indicated that this phenomenon was more widespread in the Andes than documented by Dalmayrac and Molnar (1981). Coney and Harms (1984) ascribed the onset of Basin and Range extension in the Cenozoic to such gravitational stresses generated by a thickened crust in an Andean setting over a thinned lithosphere. The extensional structures described by Dalmayrac and Molnar (1981) seem to have at least an order of magnitude less extension on them than the rifts of the Basin and Range province and may be related to left-lateral strike-slip faulting along the Peruvian Andes (cf., Dewey and Lamb, 1991, Fig. 9 and Mercier et al., 1991, Fig. 3) rather than to gravity spreading.

k42) Rifts associated with zones of continental collision. Three different environments of rifting are associated with the collision of continents: 1) *Lines of extension* that radiate from points at which collision commences; 2) *regions of extension* abutting against sutures, and 3) *nodes of extension* in areas of complex deformation in fore- and hinterlands shattered by collisions. Impactogens (k421), rifts forming along intracontinental convergence belts (k422), and pack-ice-type rifts (k423) correspond with these three environments, respectively.

k421) Impactogens. Impactogens are rifts that originate as a result of tensional stresses set up in a continent when it is hit by a pointed promontory of another continent (Şengör, 1976b, 1987b; Şengör et al., 1978). The best example today is the Upper Rhine graben between Germany and France (Fig. 2.9), which formed in the medial Eocene upon collision in the Alps (Şengör, 1976a,b, 1987b; Şengör et al., 1978). Impactogens are commonly solitary rifts, but several impactogens may form along a long front of collision, if more than one promontory collide with the opposing continent (e.g., Oslo Graben and the Viking/Central Graben in the North Sea along the Variscan collision front in northern Europe (Şengör, 1976b; see Fig. 2.6B).

k422) Rifts forming along intracontinental convergence belts. These rifts are similar in principle to those described under k413 (rifts associated with compressional arcs) and indicate the elongation of the orogen along its trend during post-collisional convergence. The north-south grabens in southern and central Tibet (Fig. 2.2), which formed as a consequence of the shortening and east-west elongation of the Tibetan high plateau following collision along the Indus-Yarlung suture represent the best active examples of these (Fig. 2.18; also see Tapponnier et al., 1986, Figs. 1 and 2).

k423) Pack-ice-type rift basins. When a continental collision generates first impactogens and then rifts related to ongoing intracontinental convergence, along with conjugate strike-slip faults that help the sideways elongation of the shortening region along the orogen, the whole deformed area becomes divided into rigid and semi-rigid blocks, in central Europe termed *Schollen* (see Dewey and Şengör, 1979, footnote 1), that move with respect to one another along compressional, extensional, and strike-slip boundaries similar to drifting pack-ice. In such a setting, rifts and grabens form in diverse shapes and orientations, as best exemplified today by the *Schollen* regime of Central Europe (Fig. 2.10; e.g., Lotze, 1937, 1971).

k5) Triple-Junction Rifts

Triple-junction rifts form at or near unstable triple junctions, at which plate evolution dictates the generation of "holes" owing to failure to create subduction zones along a plate boundary, commonly because one or more plates meeting at the triple junction consist of buoyant lithosphere. Şengör (1979a) pointed out that the FFT junction at the meeting point of the North and the East Anatolian faults at Karlıova in eastern Turkey had to generate the complex Karliova extensional basin because the T (i.e., "trench") arm of the triple junction had to function as a broad zone of convergent high strain instead of a line of surface consumption. Şengör et al. (1980, 1985) presented another example at the Bitlis suture/East Anatolian fault/Dead Sea fault junction in southeastern Turkey (the former Kahramanmaraş triple junction), which led to the formation of the Hatay rift and the Adana basin in the later Miocene following the replacement of a trench with a suture zone along the Bitlis thrust belt. Ingersoll (1982a) interpreted the complex extensional tectonics of the western United States and Mexico, involving the Basin and Range province, Rio Grande rift, and the Gulf of California transtensional system, in terms of the non-rigid response of the North American plate to the instability of the Mendocino triple junction.

Dynamic (Genetic) Classification of Rifts (Fig. 2.5)

Rifts also may be classified according to the origin of forces that lead to rifting. Şengör and Burke (1978) proposed that stresses that cause rifting may be

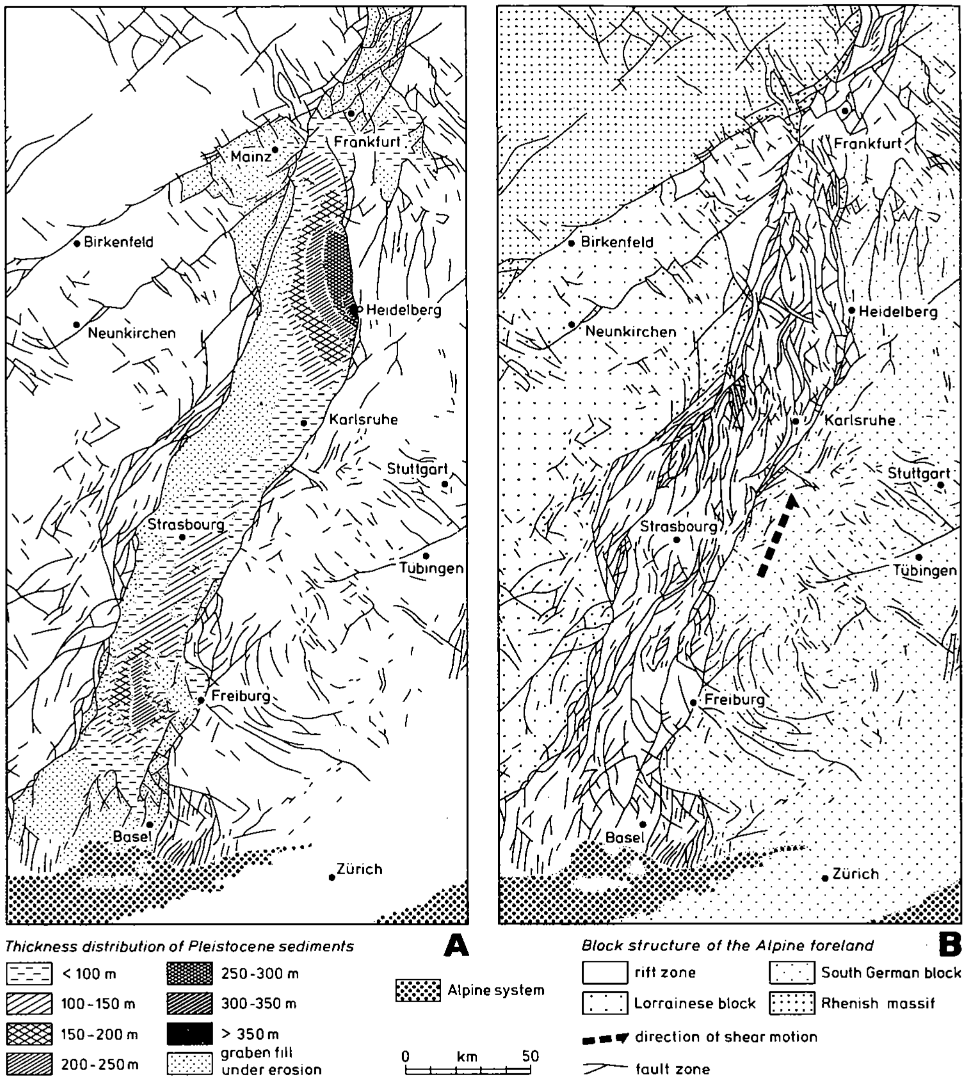
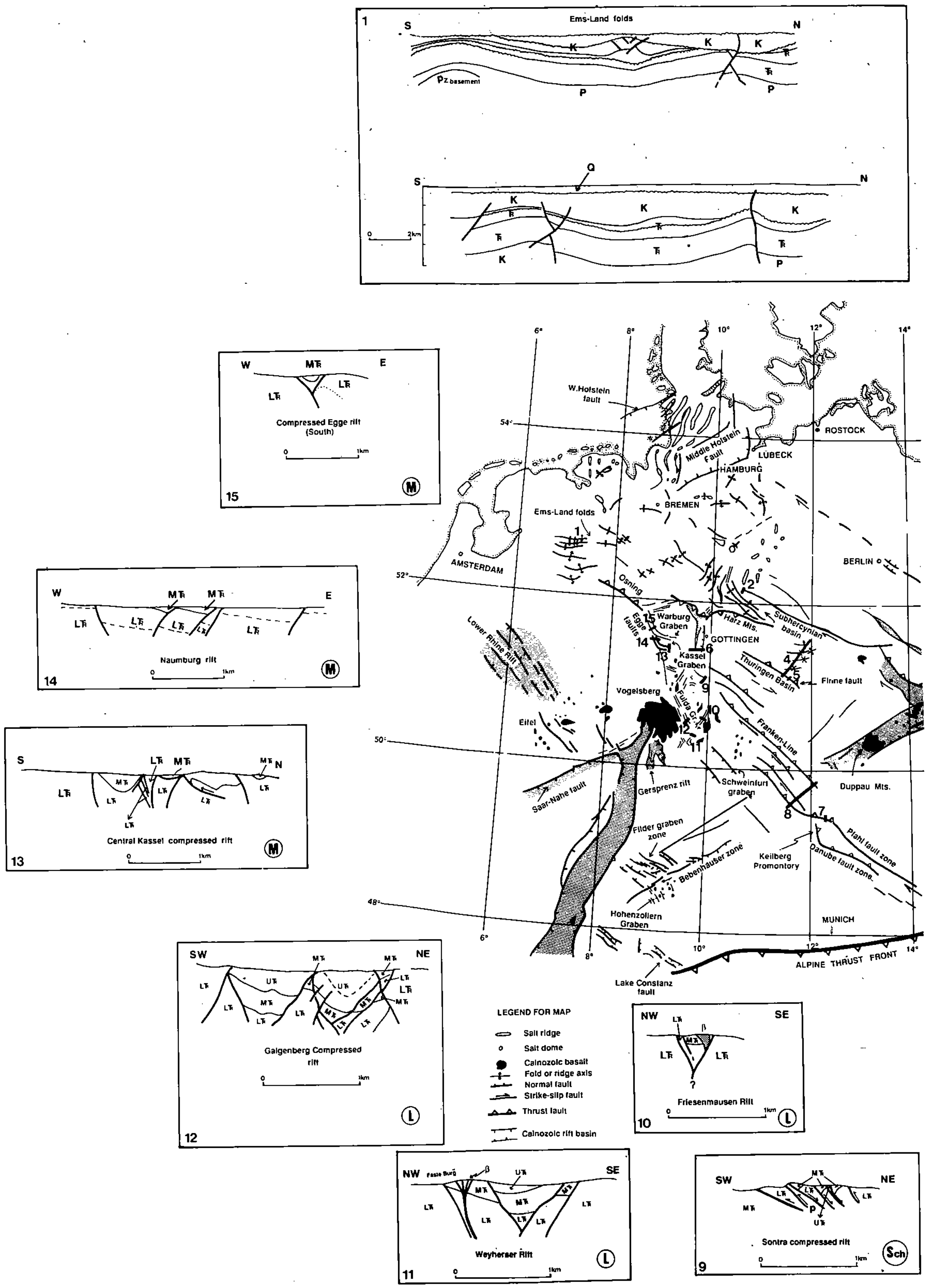


Figure 2.9 The type impactogen, the Upper Rhine rift.
A. Distribution of Pleistocene sediments outline modern Upper Rhine rift, and their thicknesses give an idea of activity of rift during Quaternary. (Reproduced with permission from Illies and Greiner, 1978, Fig. 7.)
B. Modern Rhine rift is functioning as a sinistral strike-slip zone with local shortening in its middle segment, and extension in south and in north. This map illustrates distribution of Quaternary faults. (Reproduced with permission from Illies and Greiner, 1978, Fig. 7.)

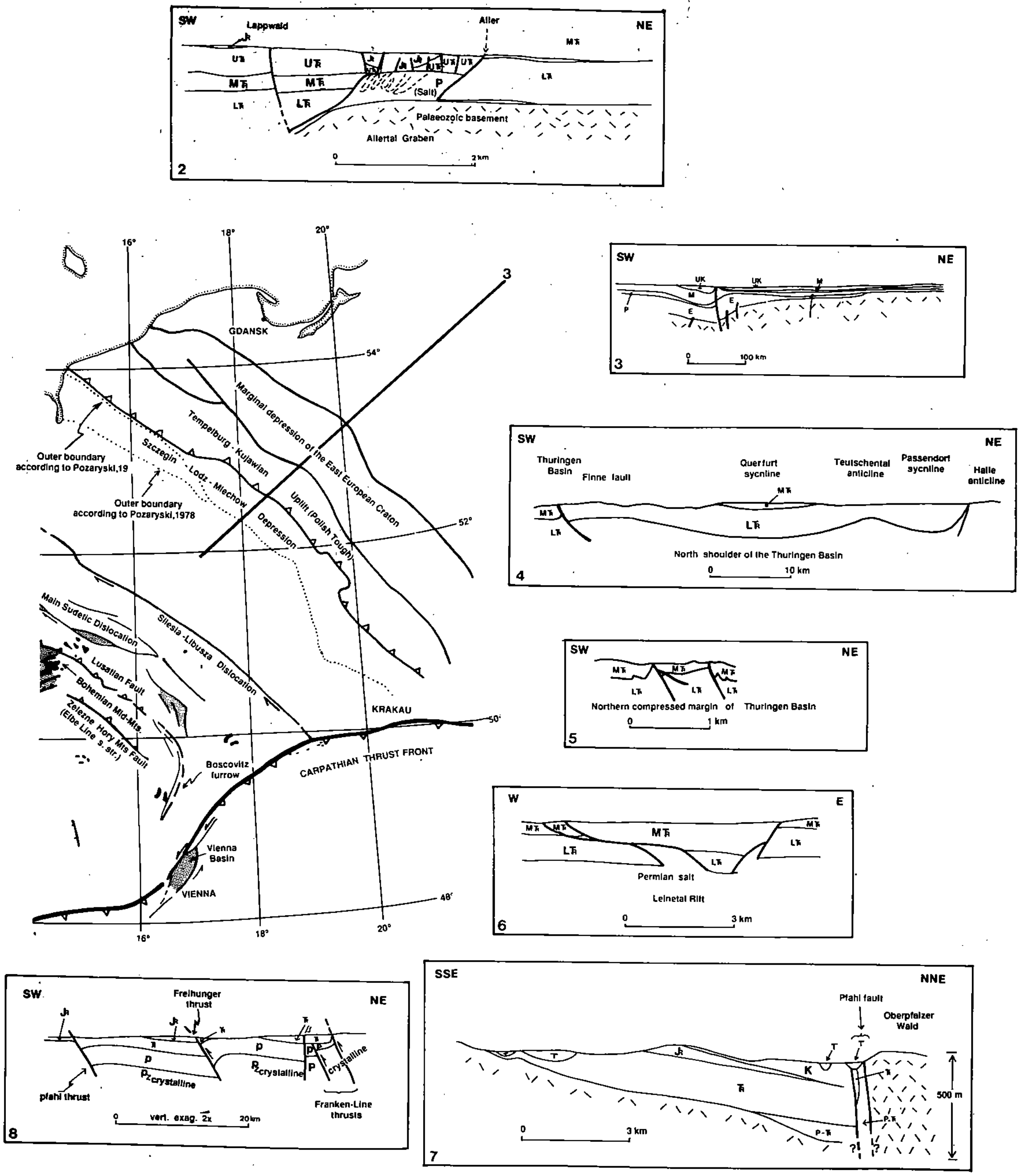
imposed on the lithosphere directly by the mantle beneath it or they may result from two-dimensional plate evolution. Accordingly, they termed these two modes of rifting *active* and *passive*. Although this subdivision has since been criticized by some (e.g., White and McKenzie, 1989), the criticism neglects the time dimension and thus is void. Moreover, this classification has proved very useful in understanding the underlying mechanisms of the multifarious styles and settings of rifting (e.g., see the contributions in Morgan and Baker, 1983a; also see Kazmin, 1987). In regional-geological (e.g., Ingersoll et al., 1990) and tectonic-modeling (e.g., Ingersoll, 1982a) studies, it is also helpful. I shall accordingly use it here also. The proposal to call passive rifting "closed-system rifting" and active rifting "open-system rifting" by Gans (1987) is misleading because it is not necessarily obvious with respect to which parameter the system is considered open or closed (crustal addition, geochemical reservoir tapped, plate boundary network, or original sedimentary provenance). Consequently, it is avoided in the following paragraphs.

d1) Active rifting

"Active rifting" is rifting caused by mantle upwelling associated with hot spots in the mantle (Burke and



1
Ems-Land folds
S
N
K
P
Pz basement
Q
0 2km
15
W
E
Compressed Egge rift (South)
0 1km
14
Naumburg rift
13
Central Kassel compressed rift
12
SW
NE
Galgenberg Compressed rift
11
NW
SE
Weyherser Rift
10
Friesenmausen Rift
9
Sontra compressed rift
W.Holstein fault
Middle Holstein Fault
ROSTOCK
LÜBECK
HAMBURG
BREMEN
Ems-Land folds
AMSTERDAM
BERLIN
Osning
Egge faults
Warburg Graben
Harz Mts.
Subhercynian basin
GOTTINGEN
Kassel Graben
Thuringen Basin
Finne fault
Lower Rhine Rift
Vogelsberg
Fulda Gr.
Eifel
Franken-Line
Saar-Nahe fault
Gersprenz rift
Schweinfurt graben
Duppau Mts.
Filder graben zone
Bebenhauser zone
Keilberg Promontory
Pfahl fault zone
Danube fault zone
Hohenzollern Graben
MUNICH
ALPINE THRUST FRONT
Lake Constanz fault
LEGEND FOR MAP
Salt ridge
Salt dome
Cainozoic basalt
Fold or ridge axis
Normal fault
Strike-slip fault
Thrust fault
Cainozoic rift basin



2
SW
NE
Lappwald
Aller
(Salt)
Palaeozoic basement
Allertal Graben
0
2 km
3
SW
NE
0
100 km
Outer boundary according to Pozaryski,19
Outer boundary according to Pozaryski,1978
GDANSK
Marginal depression of the East European Craton
Tempelburg - Kujawian
Uplift (Polish Tough)
Szczegin - Lodz - Miechow
Depression
Main Sudetic Dislocation
Silesia-Libusza Dislocation
Lusatian Fault
Bohemian Mid-Mts.
Zelezne Hory Mts Fault
(Elbe Line s. str.)
KRAKAU
CARPATHIAN THRUST FRONT
Boscovitz furrow
Vienna Basin
VIENNA
4
SW
NE
Thuringen Basin
Finne fault
Querfurt sycnline
Teutschental anticline
Passendorf sycnline
Halle anticline
North shoulder of the Thuringen Basin
0
10 km
5
SW
NE
Northern compressed margin of Thuringen Basin
0
1 km
6
W
E
Permian salt
Leinetal Rift
0
3 km
8
SW
NE
Freihunger thrust
pfahl thrust
crystalline
crystalline
crystalline
Franken-Line thrusts
0
vert. exag. 2x
20 km
7
SSE
NNE
Pfahl fault
Oberpfalzer Wald
500 m
0
3 km

Figure 2.10 Cenozoic pack-ice-type rifts and related structures in central Europe. Stippled areas are Eocene to Recent rifts. All structures shown were active at some time during the Cenozoic, although their ancestry is variable. As a rule, most NW-SE-striking fault systems originated during the late Paleozoic, probably as right-lateral strike-slip faults (see Arthaud and Matte, 1977) and were reactivated in the Mesozoic, mainly as right-lateral strike-slip faults, until they all became transpressional structures in Late Cretaceous to the present, nucleating many pull-apart basins (e.g., those along the Main Sudetic dislocation: see Illies and Greiner, 1976). The NNE-trending basins mostly formed in Late Jurassic (in northern Germany) and in early Cenozoic (in south Germany and France) as extensional structures; and most remained extensional to the present (with exceptions, such as the compressed Egge rift).

Many features played more than one, not uncommonly contradictory role during their history, even within the Cenozoic, as exemplified most dramatically by Miocene and younger thrusting along the central part of the eastern master fault of the Upper Rhine rift. Alpine shortening has further accentuated the disjointed nature of the foreland and also led to alkalic mafic volcanicity.

The map was compiled from numerous sources, but notably from: Carlé (1950), Knetsch (1963), Svoboda (1966), Lotze (1971), Illies and Greiner (1976), Pozaryski (1977), and Meiburg (1982). Cross sections were taken from Schröder (Sch: 1925: Sontra rift); Lemcke (L: 1937: Friesenmausen, Galgenberg, and Weyherser rifts); Martini (M: 1937: Kassel, Naumburg, and Egge rifts); (Lotze, 1971: all the rest).

Dewey, 1973; Burke and Whiteman, 1973; Dewey and Burke, 1974; Thiessen et al, 1979; Morgan, 1981; White and McKenzie, 1989; Hill et al., 1992; also, see Sleep, 1987, and Davies and Richards, 1992). In such environments, rifting is thought to result from the tension created by the extrados stretching caused by doming (Cloos, 1939). In fact, Cloos (1939, p. 428–435) has shown that in a dome of 600 km diameter uplifting a brittle layer of about 12 km thickness to some 4 km, a stretching of some 4 km may be obtained. His experiments with wet clay cakes have further shown that, of this stretching, only 1/4 or even 1/5 was expressed in the faults that formed. His conclusion was that doming was enough to *initiate* stretching and graben building.

Studies since Cloos (1939) have shown that although doming is sufficient to *initiate normal faulting and graben building*, it is not sufficient to *maintain rifting* and to create anything like our present rift valleys in Africa or the Rhine Graben. Indeed, Illies (1967) found, during a palinspastic restoration of the Rhine Graben extension, that a minimum of 4.8 km of extension remained unaccounted for after he had eliminated all effects of the doming. This figure, Illies (1967) ascribed to actual lithospheric extension.

Two views have been advanced to explain the origin of the extension not related to extrados extension of domes rising above hot-spot jets. One ascribes the rifting to basal shear stresses induced by a spreading plume head beneath a dome (e.g., Sorochtin, 1974; Neugebauer, 1983). The other holds the potential energy of the rising dome responsible for driving the rifting (see Richter and McKenzie, 1978, section 2; White and McKenzie, 1989). All of these factors probably contribute to maintaining the active rifting process at its habitually slow pace of less than 1 cm/y. Left to their own means, most active rifts quit eventually before leading to ocean generation (i.e., to thalassogeny). For a rift to grow into an ocean, extension along it must become compatible with, and required by, the evolving world-wide plate mosaic, whence extension rate grows above 1 cm/y. In the active mode of rifting, this generally happens when several hot spots are aligned along a line favorable to the generation of a future spreading center as Dewey and Burke (1974) and Crough (1983) have pointed out.

d2) Passive rifting

"Passive rifting" refers to a mode of rifting in which the mantle under the rifting area plays only a passive role. In the passive-rifting mode, extension is caused by two-dimensional motions of lithospheric plates. In this mode of rifting, there is no pre-rift doming related to a hot spot (Şengör and Burke, 1978). Kinematic mechanisms reviewed above under the headings k22, k31, k32, k33, k411, k412, k413, k421, k422, k423 and k5 all may form rifts in a "passive-rifting mode."

In the following section, I review the tectonic-sedimentation interplay in three types of fossil rifts only, using both fossil and modern examples of their particular classes. These are: 1) aulacogens, 2) collision rifts, and 3) intracontinental wrench basins. These three types of rifts constitute by far the commonest members of the terrestrial repertoire of fossil rifts, apart from the ones preserved along the Atlantic-type continental margins. This review uses the classification offered in this section as a framework because aulacogens, collision-related rifts, and intracontinental wrench-related extensional basins have numerous kinds and evolve through many stages, illustrating the categories of rifts discussed above. Figure 2.5 can be used as a "flow chart" to follow the evolutionary histories of the various kinds of rift basins reviewed in the following sections.

Aulacogens

Brief History of the Term

The term *aulacogen* was introduced into the tectonic terminology in an originally unpublished lecture by Shatsky, delivered on May 21, 1960 (Shatsky, 1964).

In Shatsky's usage, the term *aulacogen* was intended to designate narrow, elongate, and fairly straight depressions striking into cratons, commonly from reentrants facing an adjoining larger basin or a mountain belt. Shatsky had recognized these structures as a distinct group of cratonic structures in the 1940s and called them "transverse basins and transverse fractures," considering them a subset of his "marginal transverse structures of old platforms" (Shatsky, 1946, p. 57; Schatski, 1961, p. 99–100). Shatsky noted that they appeared genetically related to the larger basin into which they opened and that they internally segmented cratons. Because Shatsky came to recognize the aulacogens in the course of his studies on the Riphean evolution of the East European platform (i.e., shortly after the craton was consolidated), he considered them among the most important structural elements of the *early stages of the development of the craton as craton.* He noted that they localized zones of maximum thickness of the Riphean and the lower Paleozoic sedimentary sections and volcanic rocks. Shatsky observed that the lower parts of the aulacogen fills corresponded with the weakly metamorphosed and "peculiarly folded" equivalents of the adjacent miogeosynclinal rocks!

Shatsky's recognition of aulacogens was not confined to the East European platform (cf., Shatsky, 1964, p. 544). He compared the Great Donbass basin with the Wichita basin in the United States (Shatsky, 1946; Schatski, 1961, p. 81–119) and considered aulacogens common structures on cratons.

Shatsky was a convinced fixist and interpreted the origin of aulacogens in the spirit of fixism. He believed that the cause of the subsidence that generated both the aulacogen and the larger basin into which it opened was a density increase in the lower crust and/or in the upper mantle. He held a pre-existing planetary network of fractures responsible for the geometry of the aulacogens. Owing to this predetermined location and orientation, aulacogens were subjected to repeated rejuvenations (what Shatsky called "posthumous movements"), which in some cases led to a weak folding and some metamorphism of the sedimentary fill of the aulacogen.

Following Shatsky's initial recognition, aulacogens were studied extensively and classified according to various criteria: Bogdanov (1962) and Bogdanov et al. (1963), for example, distinguished "early" and "late" aulacogens, the former corresponding with those that form early in the consolidation of a craton (e.g., the Pachelma aulacogen in the East European platform: Fig. 2.2; see Klubov and Klevtsova, 1981) and the latter with those that form substantially later than the formation of the craton (e.g., the Donetz basin in the East European platform: Fig. 2.2; see Tchirvinskaya, 1981). Bogdanov (1962) also classified aulacogens according to their positions in a craton: "intra-platformal, transverse, and longitudinal!" Later, Novikova (1964) emphasized that the aulacogens of the East European platform predated its syneclises.[3] Even some orogenic belts came to be intrepreted as "aulacogens of an extreme case" (e.g., the Pyrenees and the Greater Caucasus: von Gaertner, 1969)! Despite the flurry of activity on aulacogens after Shatsky's death (for recent Soviet reviews of aulacogens, see Milanovsky, 1981, 1983, 1987; Khain and Michailov, 1985, esp. p. 150ff; p. 144–147 of the German translation), not much progress was made in understanding the origin of these structures until the rise of plate tectonics.

Origin of Aulacogens

Shatsky's explanation for the origin of the aulacogens was not only ad hoc, but also an unfalsifiable existential statement, which probably explains why not much progress occurred in understanding the origin of these structures after his death. By contrast, aulacogens have been interpreted in the framework of the theory of plate tectonics in terms of testable hypotheses with considerable success (for general reviews, see Burke, 1976, 1977a, 1980; Milanovsky, 1981; Mohr, 1982; Şengör, 1987a).

No apology is needed in returning to Shatsky's original definition of aulacogens, not only because he invented the term, but also because his definition is so clear and useful. The following is a summary, compiled from several of his publications (esp., Shatsky, 1964): An aulacogen is a narrow, elongate and fairly straight depression trending into a craton commonly from a reentrant adjoining a major basin.[4] An aulacogen commonly contains a sedimentary section at least three times as thick (commonly 10–12 km: Khain and Michailov, 1985) as the section found on the surrounding craton (commonly betwen 1 and 5 km: Khain and Michailov, 1985) into which it trends and which it segments. This sedimentary section generally correlates with the even thicker sedimentary content of the larger basin toward which the aulacogen opens. Aulacogens may be simple or complex. Complex aulacogens contain several troughs separated by internal uplifts that may be mini-foldbelts.

Figure 2.11 illustrates and summarizes the most significant and the best-documented processes that are believed to lead to the formation of aulacogens. In the following discussion of these mechanisms, I illustrate the common sedimentary environments that form in aulacogens and their products by way of the specific examples I present.

1. Doming-rifting-drifting hypothesis (path d1–k2–k21–g3 in Figure 2.5: products commonly afrotype).

Figures 2.11A, C, D, and 2.12 illustrate the "doming-rifting-drifting" hypothesis of the origin of aulacogens. In the framework of the classification given above, this hypothesis (Burke and Wilson, 1972; Burke and Dewey, 1973; Burke and Whiteman, 1973) interprets the origin of aulacogens as rifts that form through the active participation of a mantle plume, creating a dome with common diameters ranging from hundreds of kilometers (Burke and Whiteman, 1973) to about 2000 km (White and McKenzie, 1989; for this size range, see also Kazmin, 1987). Thiessen et al. (1979) found that in the late Cenozoic history of Africa, hot-spot separations vary between 2 to 15 degrees (i.e., about 220 to 1650 km) with uplift diameters ranging between similar magnitudes. These observations agree well with the cell sizes obtained during convection experiments undertaken by Richter and Parsons (1975), assuming a depth of upper-mantle convection of about 700 km (for numerical modeling reaching similar conclusions, see White and McKenzie, 1989). (For the hypothesis of hot spots generated by whole-mantle convection, see Morgan, 1981 and Davies and Richards, 1992.) According to the doming-rifting-drifting model, taphrogeny and the consequent thalassogeny (or koilogeny) is preceded by the formation of several such domes rising to structural mean heights ranging from 1 to above 2 km (see Cloos, 1939; White and McKenzie, 1989), that may be expressed at the surface as topographic domes rising over 2 km, with peaks surpassing 4 km, as in Ethiopia.

It may be possible to document and date such uplifts in the geological record, long after their topographic expression has vanished, by means of the peripheral clastic wedges they shed. For example, the Miocene clastics of western Kenya and Uganda were produced by accelerated erosion during arching over the sites of the future western and eastern rifts and were later preserved beneath the younger volcanics (e.g., King, 1970). Figure 2.13 documents a similar situation from Patagonia, where Late Triassic to Early Cretaceous rifting of the Salado, Colorado, San Jorge, and other smaller rifts, which later led to opening of the South Atlantic ocean (cf., Burke, 1976), was preceded by uplift and erosion. Depositional products are today seen preserved in such fluviatile clastics as those of the Tralcan and those correlative with the upper part of the Puesto Piris porphyries, all of which are now blanketed by widespread Jurassic intermediate volcanic rocks in the region. Renewed doming was recorded by the Los Adobes, Las Heras, Cerro Barcino and La Rueda clastics partly on the Somoncura Massif, which are, in part, correlative with the rift fill (e.g., Colorado interior: Fig. 2.13C: see esp. Uliana and Biddle, 1987; Uliana et al., 1989). Such peripheral clastic wedges shed by plume-generated uplifts have been recognized even in cases where they were later involved in major orogeny! For instance, the pre-

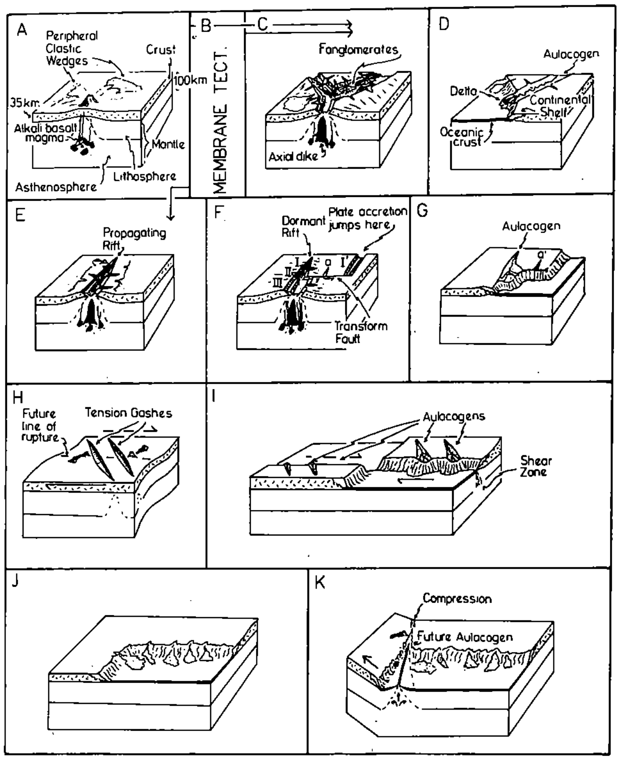


Figure 2.11 Schematic block diagrams illustrating most significant and best documented plate-tectonic processes that lead to formation of aulacogens (reproduced with permission from Şengör, 1987a, fig. 1). **A-C-D**: Doming-rifting-drifting hypothesis (see also fig. 2.13); **B-C-D**: Hypothesis of aulacogen formation through "membrane stresses" (see also fig. 2.16); **E-F-G**: Rift-tip-abandonment hypothesis; **H-I**: Strike-slip-related secondary-extension hypothesis; **J-K**: Continental-rotation hypothesis. See text for discussion and other hypotheses.

Varanger shallow-marine sandstones and quartzites (late Riphean in age: 800–750 Ma) of the "Tosasfjaellet basin" (now found in the Saerv Nappe of the Middle Allochthon of the Scandinavian Caledonides) were interpreted to have been shed from a plume-generated uplift, whose rifting later led to the birth of the Iapetus ocean (Kumpulainen and Nystuen, 1985, esp. Fig. 4).

Such plume-generated domes are, in places, sites of alkaline to peralkaline magmatism (see Thiessen et al., 1979, Fig. 3 for the distinction between high spots (i.e., those uplifts without magmatism) and hot spots (i.e., those with magmatism). The volcanic products of this magmatism preceding rifting are commonly preserved, as are the clastic sediments shed from the uplifts, usually in the form of volcaniclastics, more

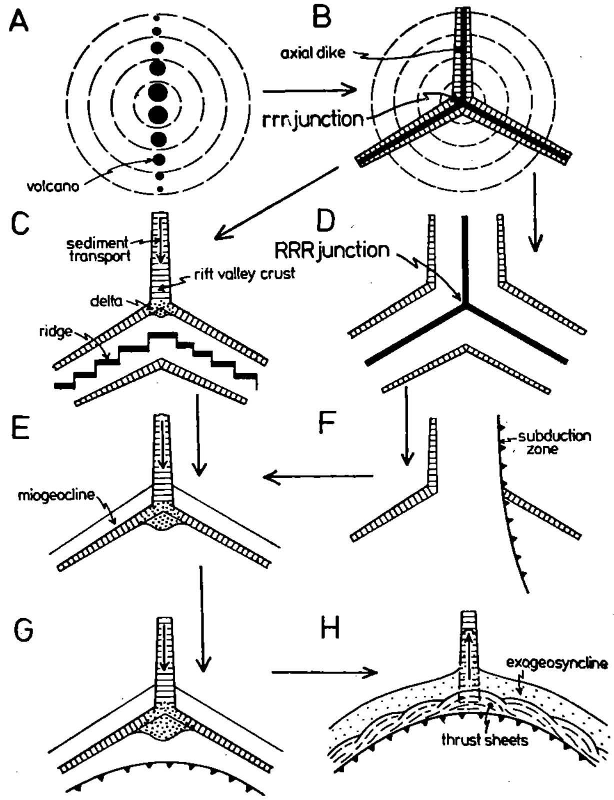


Figure 2.12 Schematic origin and evolution of mantle-plume-generated triple junctions, leading to doming-rifting-drifting sequence creating oceans and aulacogens (from Burke and Dewey, 1973, Fig. 2, reproduced with permission). **A.** Uplift develops over plume with crestal alkalic volcanoes; **B.** Three rift valleys develop at an rrr (rift-rift-rift) junction (e.g., Nakuru: see Fig. 2.6A); **C.** Two rift arms develop into a single plate margin (ridge) and continental separation ensues, leaving third rift arm as an aulacogen, down which a major river may flow and at the mouth of which a major delta may develop (e.g., Limpopo: see Fig. 2.2); **D.** Three rift arms develop into spreading centers meeting at an RRR (ridge-ridge-ridge) junction (e.g., Red Sea-Gulf of Aden-Wonji fault belt in Ethiopian rift system: Mohr, 1971; also see Burke and Dewey, 1973, Fig. 3A); **E.** Atlantic-type continental margin evolves with growth of delta at mouth of aulacogen and continental shelves (e.g., Mississippi: see Fig. 2.2); **F.** One arm of RRR system begins to close by subduction. If ocean is sufficiently wide (>1000 km), a magmatic arc will develop along its margin and any sediments in closing arm will be deformed (e.g., Lower Benue Trough); **G.** Atlantic-type continental margin with continental shelves and aulacogens approaches a subduction zone (e.g., present Sirte rift approaching the Hellenic subduction zone); **H.** Continental margin collides with subduction zone, collisional orogeny ensues, sediment transport in the aulacogen reverses polarity, and aulacogen may become tectonically rejuvenated (e.g., Athapuscow: Fig. 2.2)

rarely rhyolite flows or welded tuff (e.g., in Patagonia; see Uliana and Biddle, 1987; Fig. 2.13). In former domes, the volcanics crowning the dome are now

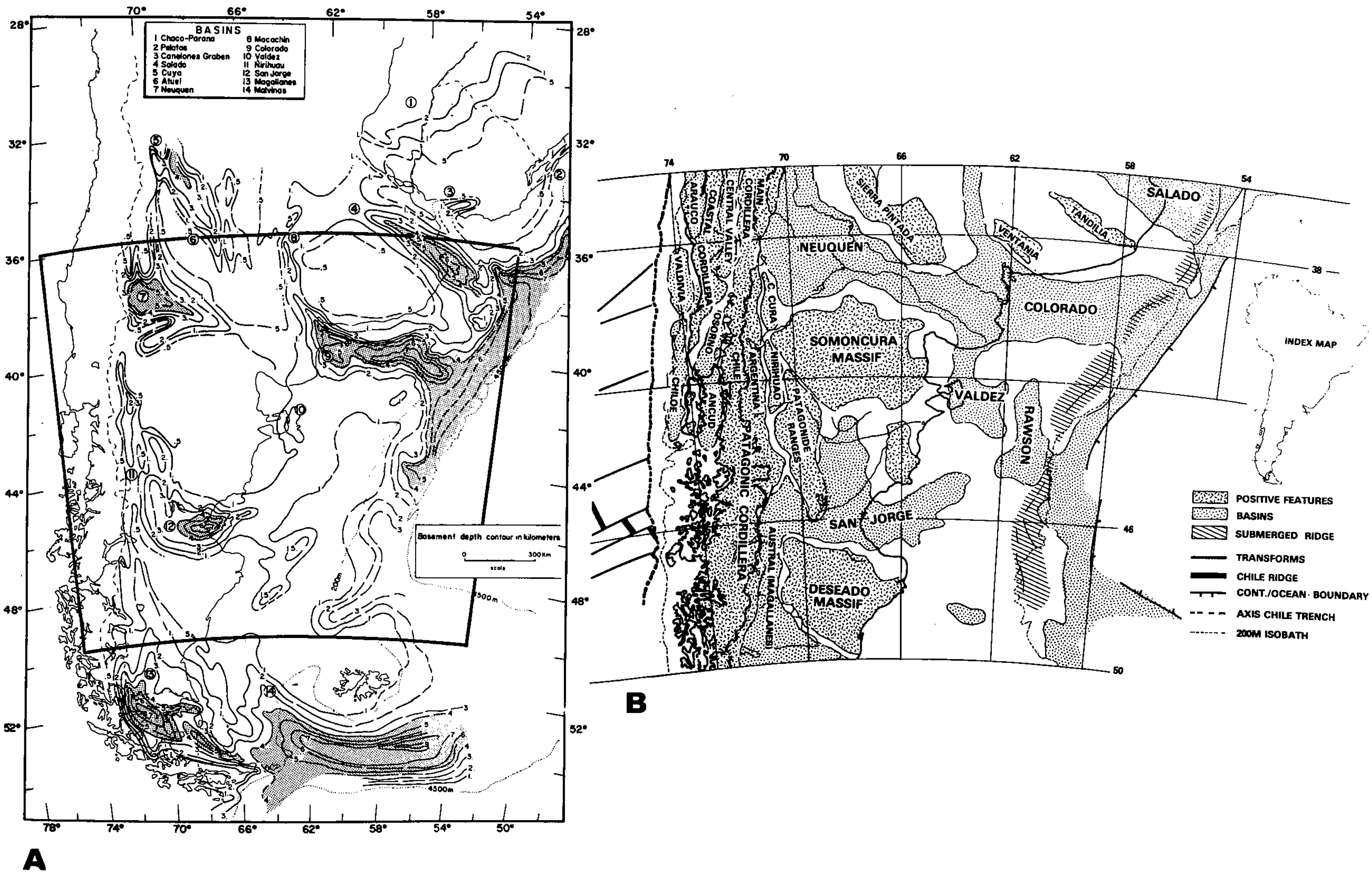


Figure 2.13 Mainly Early Cretaceous aulacogens of Argentine Atlantic continental margin (for a graphic history of their development see Uliana et al., 1989, Figs. 2,6,8,10, and 12).
A. "Technical" basement topography in Argentina, obtained by mapping all seismic velocities higher than 4.5 km/sec. Regions at depths greater than 4 km are stippled, and among them two prominent aulacogens, Salado and Colorado stand out (in Fig. 2.13B, they lie on both sides of the label "Argentina"). (Reproduced with permission from Urien and Zambrano, 1973, Fig. 4.)
B. Location map showing geological provinces, whose Mesozoic stratigraphy is summarized in fig. 2.13C. (Reproduced with permission from Uliana and Biddle, 1987, fig. 1.)

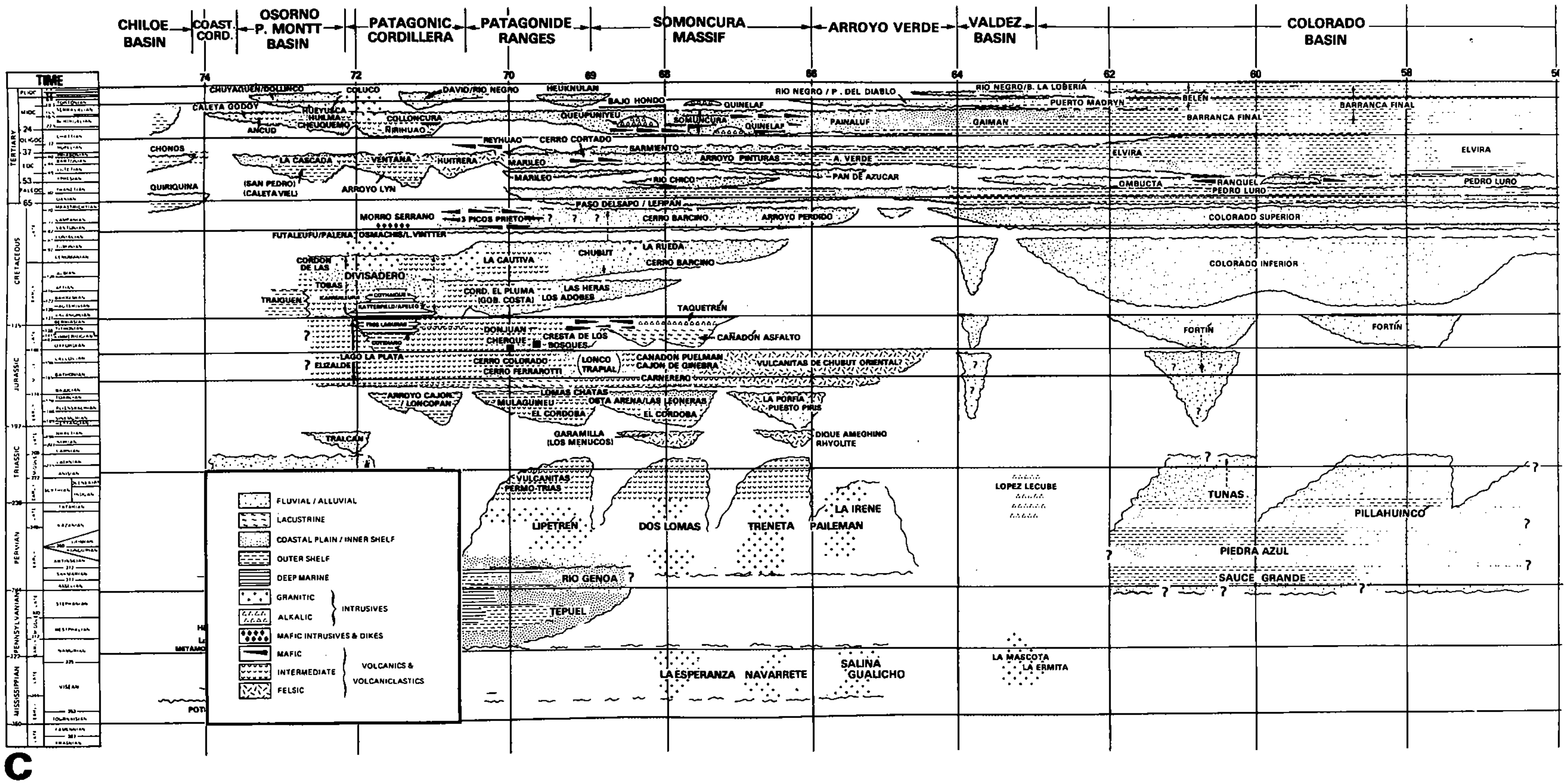


Figure 2.13 C. Chronostratigraphy of a strip on Fig. 2.13B, showing Late Triassic-Jurassic uplift at future site of Colorado rift (by the absence of corresponding deposition) and correlative erosional products represented by clastic rocks of Tralcan, those under the Puesto Piris porphyries, and those of the Osta Arena and Las Leoneras units to the south and west. (Reproduced with permission from Uliana and Biddle, 1987, plate 1.)

commonly eroded, but their former presence can be surmised from the preserved feeder dikes or other intrusions into the basement of the dome. One of the best places where such intrusions are today seen in the cores of former uplifts is Nigeria. There, the "younger granite magmatism" (cf., Jacobson et al., 1958; Whiteman, 1982) has produced igneous rocks ranging from alkalic granites and syenites to gabbros and basalts exposed as plutons and dikes (the ring complexes). In some cases, normal faulting accompanied granite intrusions, as a consequence of which, the rhyolitic carapaces are preserved (Jacobson et al., 1958). These intrusive rocks in Nigeria occur mostly on the northern shoulder of the Benue Trough (see Whiteman, 1982, Fig. 16; Benkhelil et al., 1988, Fig. 32–6).

The doming phase is succeeded by crestal rifting, and the initial dome-related normal faults commonly radiate away from the center of the dome in the form of a rift star (Figs. 2.11B, 2.12B; also compare Fig. 2.7A with Fig. 2.12; Burke and Whiteman, 1973, esp. Figs. 4 and 9). Cloos (1939), and following him, Burke and Dewey (1973) suggested that a triple-armed rift star with the three arms being ideally 120 degrees apart in an environment where the horizontal stress field is isotropic (similar to the ones shown in Figs. 2.6A, 2.11B, and 2.12B) may be a least-work configuration to release the dome-related stresses.This suggestion is corroborated by the many rift-rift-rift (rrr) junctions recognized on domes in Africa by Burke and Whiteman (1973).

Within these rifts, conglomerate, immature sandstone, and terrestrial mudrock, commonly representing floodplain deposits, and locally, evaporites in association with other playa facies, form. Young rifts do not possess well established drainage, and ongoing tectonism frequently creates and destroys lakes on the rift floor. Such lakes may deposit clastics, limestones, and diatomites in the Cretaceous and younger rifts, and even evaporites, with depositional sites ranging from deep lake (e.g., Lake Tanganyika, with 1.5 km water depth, the second deepest lake in the world today, is located in a rift valley along the East African rift chain: Burgess et al., 1988), to lake margin and swamp (e.g., the Middle Pliocene Awash lake in the southern Afar rift in Ethiopia: Williams et al., 1986; for well described Mesozoic lakes in fossil rifts, see Lorenz, 1988, ch. 8; Olsen, 1988). Paludal environments in rifts are usually restricted and coal forms only under appropriate climatic conditions in marshy areas temporarily shielded from coarse clastic influx (e.g., Lorenz, 1988, ch. 7). All such sedimentary deposits together form what is commonly known as the "graben facies" (Bird and Dewey, 1970) or "rift valley facies" (Burggraf and Vondra, 1982). Ash-fall tuffs (e.g., de Heinzelin, 1982, esp. Fig. 2) and Milankovich rhythmites (called "Van Houten cycles": Olsen, 1984, 1986, 1991) provide the best stratigraphic markers for intra-rift correlations because the biostratigraphic record is usually poor in such terrestrial environments. In Cenozoic rifts, the fossils of rapidly evolving vertebrates have been used for detailed biostratigraphic calibration (e.g., suids in East Africa: White and Harris, 1977), but in older rifts, one is commonly faced only with a meager palynomorph record (e.g., Asmus and Ponte, 1973), or nothing at all (e.g., Uliana and Biddle, 1987). Some ancient rifts preserve a rich vertebrate fauna associated with lacustrine invertebrates and abundant plants (e.g., the Newark Supergroup in the Newark rift: Olsen, 1988). Moreover, Seyitoğlu and Scott (1991) have recently shown the great uncertainties that may plague the usage of palynomoprphs for stratigraphic calibration in rifts in the absence of isotopic age data even in the late Cenozoic. Magnetostratigraphy is of great help in calibrating terrestrial rift fills, but in old rifts, magmatism and burial may confuse the magnetic record. So far, the most reliable and most widely used method for dating old terrestrial sedimentary rift fills is isotopic dating of the associated igneous, especially volcanic, rocks. This becomes especially powerful when combined with biostratigraphic, magnetostratigraphic and paleoclimatologic data (e.g., Witte et al., 1991; Olsen et al., in press).

In rifts, marginal fanglomerates are particularly useful in stratigraphic analysis because they record the successive levels of exposure on the rift shoulders, thereby complementing the syngraben doming record of the clastic wedges outside the rift (for active and late Cenozoic examples, see Tiercelin, 1986; Williams et al, 1986; for a spectacular fossil example, see Steel and Gloppen, 1980; for further examples from impactogens, see below). In this regard, the marginal clastics along main taphrogenic fault belts are analogous to the molasse deposits along main orogenic thrust belts, which record the uplift and unroofing history of the orogen (cf., Trümpy, 1973). Study of the marginal fanglomerates in many rifts has shown that rift-related morphogenesis and presumably related structure generation are locally episodic processes, but local episodes of relief formation (in most cases resulting from faulting) are not synchronous taphrogen-wide, or even on both sides of the same segment of a single rift (e.g., Armijo et al., 1986).

Two of the arms of a rift star (which most commonly occurs in the form of an rrr junction: Burke and Dewey, 1973; Burke and Whiteman, 1973) may eventually propagate and link with similarly or conveniently orientated arms of other rift stars to form a single divergent plate boundary, along which an ocean may open (Figs. 2.11D, 2.12C and 2.13A; Burke and Dewey, 1973; Dewey and Burke, 1974; also see Siedlecka, 1975, Fig. 7). The third, "unspread" arm will be left as an aulacogen trending into a continent from an oceanic embayment (e.g., Gulf of Guinea and the Benue Trough; the North Caspian depression and the Donetz and Pachelma aulacogens). Global plate geometry determines the lines along which a new accreting plate boundary will develop, and therefore, which arms of a group of rift stars will not produce ocean.

2. Doming-rifting hypothesis (path d1–k1–g1 in Fig. 2.5: products commonly afrotype).

Through the doming-rifting mechanism, aulacogens need not form only as failed arms of triple-(or more-)spiked rift stars. They may form as solitary rifts at continental margins as a consequence of the fortuitous location of a hot spot at or near a continental margin (see Şengör et al., 1978, Fig. 1). White and McKenzie (1989, esp. Fig. 17) illustrated a well documented example from the former (and also partly, the present) western continental margin of India, where the Reunion hot spot responsible for Deccan trap magmatism also led to the rifting of the Seychelles fragment away from India (at magnetic chron 28 time; i.e., about 65 Ma). Van Houten (1983) suggested that the Cretaceous Sirte rift of north-central Libya (Fig. 2.14A) may be a fossil example of a solitary rift that formed above a hot spot at the previously formed North African continental margin, and which is now left as an aulacogen opening into the Eastern Mediterranean.

In such solitary rifts that formed at previously existing continental margins and were later abandoned as aulacogens, the rift fill commonly becomes marine much more quickly than in those rifts that go through the doming-rifting-drifting cycle. In the Sirte rift, for example, the rift fill is exclusively marine, with the transgressive Maragh clastic unit at the base (Fig. 2.14B).

3. Rift-tip-abandonment hypothesis (path d2–k2–k22–(g2 or g3) in Fig. 2.5: products commonly afro-, more rarely aegeotype).

Figures 2.11E–G illustrate another mechanism to form aulacogens which was put forward by Grant (1971). The tip of a propagating rift[5] (Figs. 2.11E, F, segment I) may become dormant as a result of the sideways jumping of the locus of extension, forming a new rift (Fig. 2.11F, segment I′). If ocean opening then takes place along the segments I′–II–III, then the original rift tip, I, may be left as an aulacogen trending into the continent from a reentrant formed as a result of transform fault/ridge (II–III) intersection (Fig. 2.11G). The best active example is the Gulf of Suez rift (Fairhead and Stuart, 1982).

The best known aulacogen of this type is probably the Connecticut Valley rift (Fig. 2.15), widely known for its spectacular dinosaur footprints.[6] It actually consists of two basins, namely those of Hartford (in the south) and Deerfield (in the north) (Fig. 2.15A; see Eardley, 1962, p.131ff. and Figs. 9.3 and 9.4; De Boer and Clifford, 1988; Olsen, 1988; Phillips, 1988). It trends southwards into yet another rift, called the New York Bight rift (Fig. 2.15C and D; Hutchinson and Klitgord, 1988), which opens into the New York Bight, the reentrant formed by the N 40 Kelvin Lineament, an early transform margin (see Hutchinson and Klitgord, 1988, Fig. 4–2). The sedimentary fill of the Connecticut rift valley spans the time interval from the late Norian to the Toarcian inclusive, and consists entirely of terrestrial sedimentary rocks and intercalated basaltic lavas (Eardley, 1962). At the bottom of the section are nearly 3 km of relatively coarse, fluvial gray and pink arkose, conglomerate,

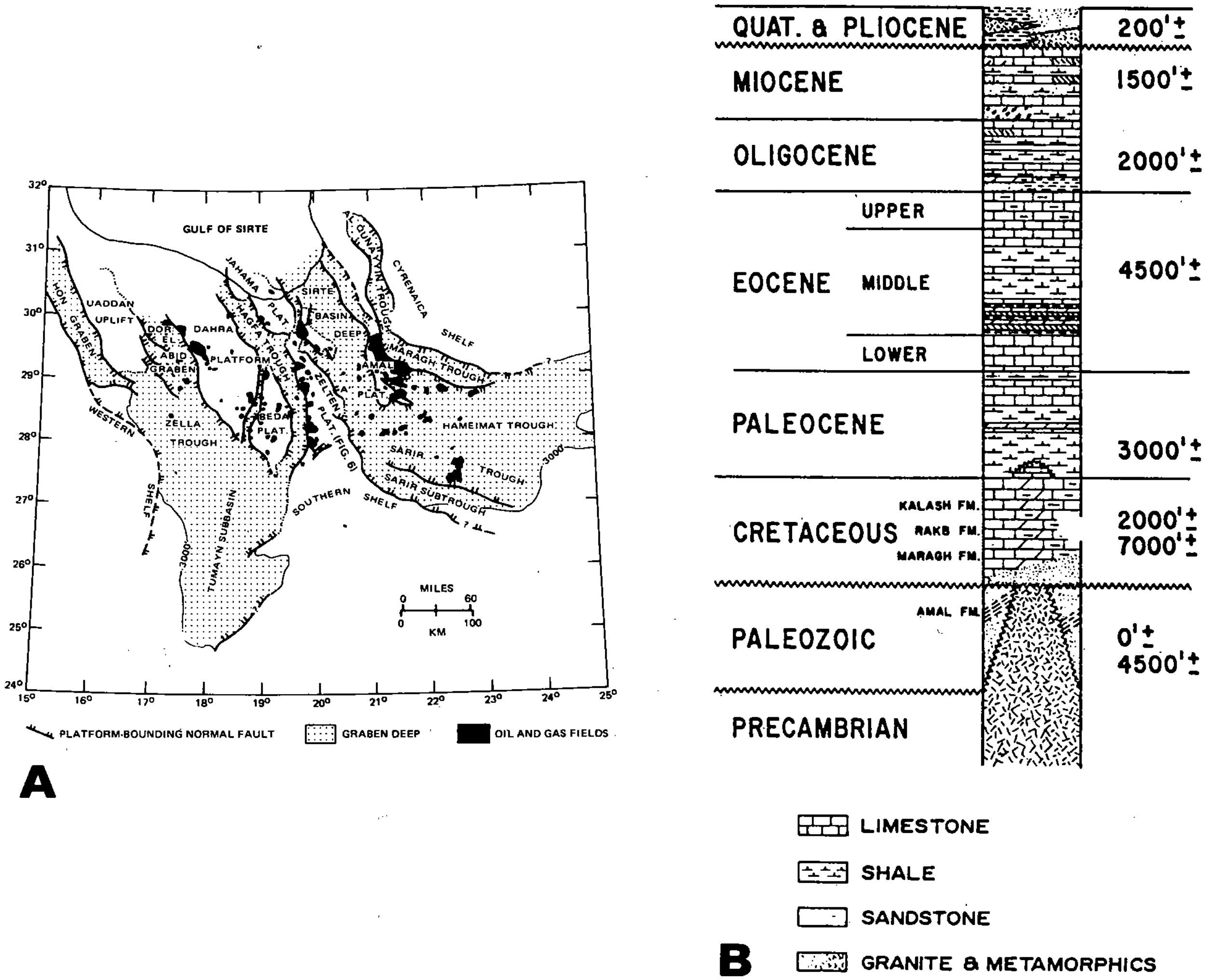


Figure 2.14 **A.** Sirte rift in Libya. Five major northwest-trending grabens have formed an unusually wide rift complex in northwest part of basin (reproduced with permission from Harding, 1983a, Fig. 5).

B. A generalized stratigraphy of the Sirte rift in Libya. Note dominance of marine section (brick pattern above Paleozoic-Cretaceous unconformity; reproduced with permission from Roberts, 1970, Fig. 7).

red feldspathic sandstone, and subordinate red siltstone and shale. These are overlain by nearly 1 km of thick, fine-grained lacustrine and paludal variegated and dark siltstone, shale, limestone, light feldspathic sandstone, subordinate coarse clastic rock, and three basaltic lava flows of Hettangian age (Ar-Ar dating: Sutter, 1988; older vintage K-Ar dates had indicated Hettangian and Sinemurian ages; see Manspeizer, 1988), which were erupted within an interval of only 6.7×10^5y, as determined by Milankovich-type lake cycles (Olsen et al. in press). In fact, Olsen et al. (in press) believe that the igneous rocks may be only 4.0×10^4y younger than the hitherto poorly constrained Triassic-Jurassic boundary.

The upper part of the section is a clastic deposit similar to the lower clastics, but is generally finer-grained. At all stratigraphic levels, the section is coarser toward the eastern master fault, along which are marginal fanglomerates called the Mount Toby conglomerate (Eardley, 1962). The total section thins from more than 5 km along the eastern margin of the rift to below 500 m in the west. As a whole, the Connecticut rift is a west-facing "half-rift", whose sedimentary fill consists of 10% conglomerate, 64%

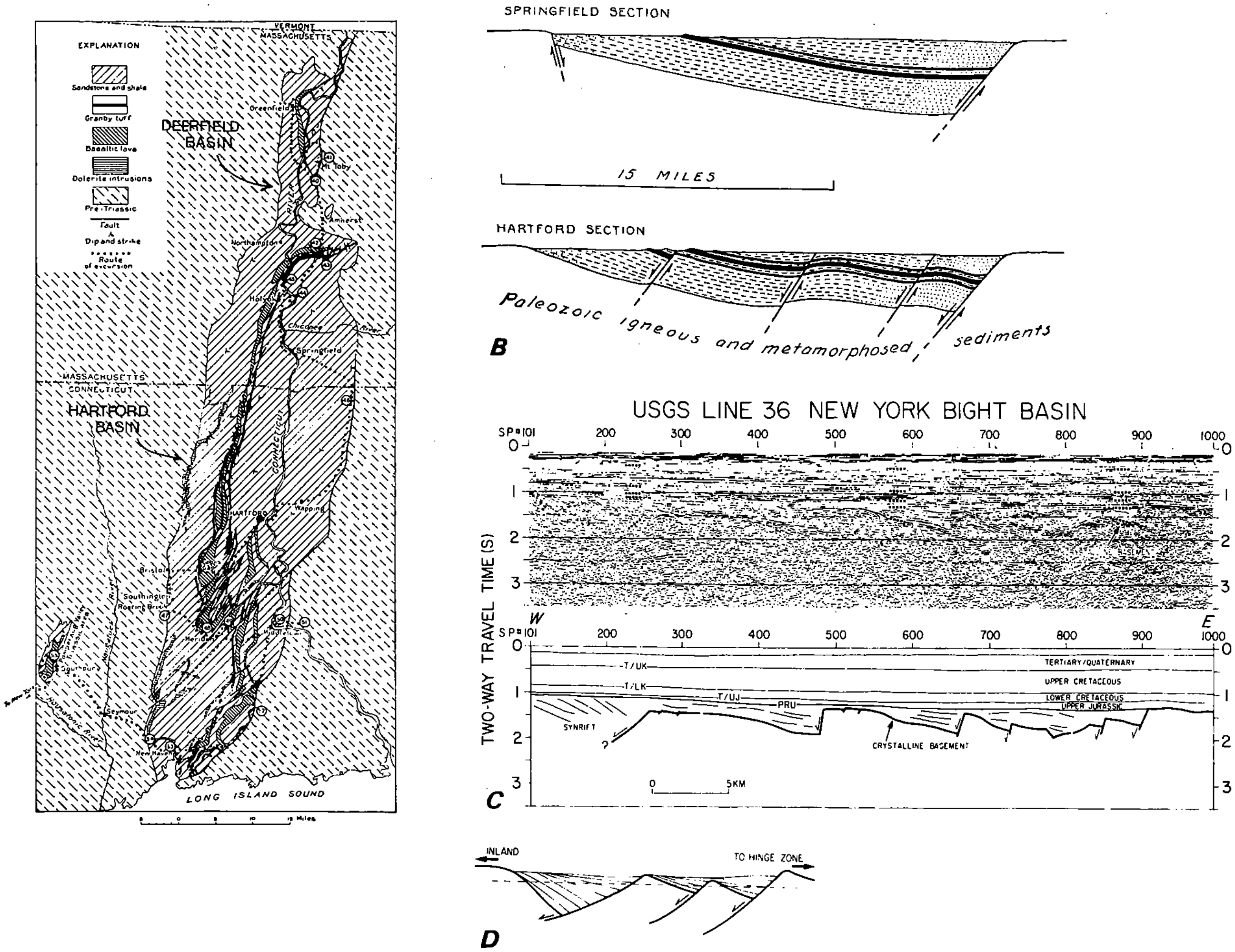


Figure 2.15 Largely onland example of an aulacogen formed by rift-tip-abandonment mechanism, Connecticut rift valley and New York Bight rift.
A. Simplified geological map of Connecticut rift valley, showing its two constituent rifts, namely Deerfield basin in the north and Hartford basin in the south. (Reproduced with permission from Eardley, 1962, Fig. 9.3.)
B. Two geological cross sections across Hartford basin, showing asymmetric nature of rift and distribution of clastic facies. See text for discussion. (Reproduced with permission from Eardley, 1962, Fig. 9.4.)
C. Submarine part of Connecticut rift valley system, New York Bight basin, likely separated from the former by a cross horst, forming Long Island (reproduced with permission from Hutchinson and Klitgord, 1988, Fig. 4-3,). Key to abbreviations: PRU: Post-rift unconformity; T/UK: top of Upper Cretaceous; T/LK: Top of Lower Cretaceous; T/UJ: Top of Upper Jurassic.
D. The asymmetric nature of New York Bight basin (compare with Connecticut rift valley; reproduced with permission from Hutchinson and Klitgord, 1988, Fig. 4-9).

sandstone, 25% shale and siltstone, and the rest limestone. In the New York Bight rift, the structure is similar to the Connecticut rift, but is more disrupted by numerous west-dipping faults, and the entire rift is unconformably overlain by an Upper Jurassic sedimentary blanket.

In older orogenic belts and/or in their forelands, such rift-tip propagation rifts are difficult to recognize. Thomas (1977, esp. Fig. 11) illustrated some possible examples related to former embayments that originated related to the ridge/transform fault intersections during the opening of the Iapetus ocean in the Appalachians.

4. Hypothesis of aulacogen formation through "membrane stresses" (path d2–k2–k22–(g2 or g3) in Fig. 2.5: products commonly afrotype).

Figure 2.11B suggests that the sequence of events shown in Figs. 2.11C–D and E–G may be initiated by extensional membrane stresses set up in a continent as a result of the change of radius of curvature of a continental plate resulting from its longitudinal motion on a non-spherical earth (Fig. 2.16; Turcotte and Oxburgh, 1973; Turcotte, 1974). As the plate moves from the equator towards the pole, its radius of curvature must become larger (i.e., the plate must become "flatter") thus increasing the length of its perimeter. This induces tension around the periphery and compression in the center as shown in Fig. 2.16. Turcotte and Oxburgh (1973) thought that this may be one way of forming marginal aulacogens that may commonly appear as solitary rifts striking from the continental plate margins into the cratons, as, they contended, in the case of the Mississippi Embayment rift. They thought that it may have been associated with crack-tip propagation following a path along the St. Lawrence Valley, Missouri, and the New Madrid tectonic zone as the North American plate moved northward. They ascribed extension across the East African Rift chain to the same stresses generated by Africa's northerly motion (Oxburgh and Turcotte, 1974). Burke and Dewey (1974) disputed the role of membrane stresses in rifting processes, however, because it seems difficult to store the elastic stresses as long as required by the longitudinal motion of plates across distances sufficient to accumulate enough stress to initate rifting.

More recently, Solomon (1987) took up membrane stresses as a possible cause for rifting, but in the context of secular cooling of the earth. He contended that, as the earth cools, it shrinks and reduces its radius of curvature. This would have the equivalent effect on overlying plates as if they had been moving from the pole to the equator (cf., Fig. 2.16) (i.e., their interiors undergo extension). Solomon thought that this may cause the formation of rift stars, some of whose arms may link to disrupt a continent and lead to ocean and aulacogen generation.

In all cases of aulacogen formation in the framework of the membrane-stresses hypothesis, pre-rift doming is not a requirement, as *k22* in their "formation path" expresses. But apart from this, it is difficult to test this hypothesis. In the interpretation of Turcotte and Oxburgh (1973), one obvious test is to check the apparent-polar-wander path of the continent in question to see whether it moved in a direc-

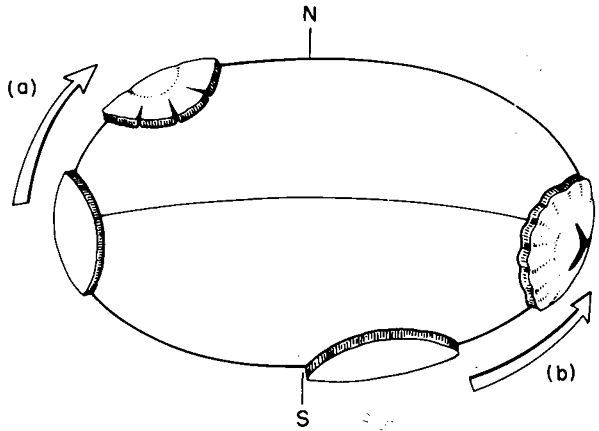


Figure 2.16 Schematic diagram illustrating mechanism of rifting through membrane stresses: a. Plate moving from equator to pole has center in compression and outer part in extension, creating radial aulacogens as solitary rifts (g1 in Fig. 2.5); b. Motion towards equator gives a central zone of extension and outer zones of compression. Central zone of extension gives rise to rifting, creating mainly rift stars (g2 in Fig. 2.5), which may propagate as plate continues to move to create rift chains (g3 in Fig. 2.5). Oxburgh and Turcotte (1974) maintained that this last configuration represents Neogene behavior of the African plate (but see Burke and Dewey, 1974).

tion expected from its rifting history. A corollary to the doming-rifting-drifting hypothesis is that the continent to rift is generally required to come to rest with respect to its underlying mantle convection pattern (cf., Burke and Wilson, 1972; Burke and Dewey, 1973). This requires that the plate move with respect to the magnetic pole during rifting only as fast as its underlying convection "pattern", which would be at most a few centimeters a year, judging from the relative motions of hot spots (Burke et al., 1973; Molnar and Franchetau, 1975; Molnar and Stock, 1987; but see Morgan, 1981). By contrast, the model of Turcotte and Oxburgh (1973) requires that the plate rift while moving fairly fast, so as not to dissipate the membrane stresses. As far as the present observations permit, the doming-rifting model for the origin of the East African Rift chain seems the more appropriate of the two, because the African plate seems to have been stationary with respect to the underying convection pattern in the mantle since the early Miocene (Burke and Wilson, 1972).

Solomon's (1987) suggestion is almost impossible to test empirically. An incomplete test would be to ascertain that all major continental assemblies in the past began rifting in their middle regardless of the direction of their motion on the surface of the earth and in the absence of any precedent doming around independent centers (hot- and high spots). This requirement is not met by the late Paleozoic Pangean disintegration (it began in the Carboniferous in the extreme east, in the present northeastern Australia).

From a sedimentological viewpoint, membrane-stress-driven rifting would ideally have no associated clastics anywhere shed from any pre-rift domes. Once the rifting commenced, the subsequent sedimentological evolution would be indistinguishable from the products of other mechanisms reviewed above.

5. Strike-slip-related secondary extension hypothesis (path d2–k3–k33–g1 in Fig. 2.5: products commonly aegeotype).

Transform-fault segments of ridge/transform intersections during the initial breakup of a continental mass may be substantial (e.g., the DeGeer line between the Barents shelf and Greenland and the Agulhas fracture zone between the Malvinas plateau and South Africa: Fig. 2.2). There will be an ever-shortening (along the strike of the transform fault) intracontinental strike-slip regime affecting the transform segments, while the separating continents are clearing each other along shear margins (cf., Emery, 1977, Fig. 2). Tension gashes and/or rotated and opened Riedel and anti-Riedel shears (cf., Wilson, 1960; Tchalenko, 1970; Harding et al., 1985, esp. Fig. 1; Sylvester, 1988) generate rifts at this stage at various angles to shear-zone strike (Fig. 2.11H). As the broad shear zone finally narrows to a throughgoing strike-slip fault zone, the previously formed rifts along the shear belt are cut by the new fault zone and left, on the resulting transform margin, as aulacogens opening into the ocean thus created (Fig. 2.11I). In this case too, pre-rift doming is not a requirement.

The Late Jurassic - Early Cretaceous rifts on- and offshore in South Africa (Figs. 2.17A,B: Rigassi and Dixon, 1972; Dingle, 1976; Tankard et al., 1982, ch. 12) and the post-Triassic rift of the Falkland Sound separating West Falkland from East Falkland (Halle, 1911; Ludwig et al., 1979) were likely formed as a result of the broad shearing along the future Agulhas fracture zone as South America began separating from South Africa.

In South Africa, the pre-rift stratigraphy is irregular because the mountains that formed during the Cape orogeny in the late Paleozoic - earliest Mesozoic stood high and their southern face sloped toward the present shoreline, which nearly coincided with the depositional limit of the post-Cape clastic Robberg Formation of probably Liassic to early Malm age (Fig. 2.17A: Rigassi and Dixon, 1972; Tankard et al., 1982, ch. 12). Wherever Robberg deposition coincided with later rifting, this formation is preserved within the rift trough (e.g., the southern part of the Plettenberg rift: Fig. 2.17A: Rigassi and Dixon, 1972, esp. Fig. 4), suggesting that there was no pre-rift doming, as expected from the postulated mode of rifting. The fact that in most South African rifts, the Robberg is not seen, is due to original nondeposition in those areas.

The rift sedimentary fill in South Africa consists of Uitenhage Group, consisting of three main units, namely, the Enon conglomerate, and the Kirkwood and Sundays River Formations (Figs. 2.17B and C; for review and literature, see Tankard et al., 1982, esp. p. 411ff). The Enon conglomerate is a coarse, red fluviatile deposit, with enormous boulders (up to 50 cubic meters!) at its base. It is interpreted to be associated with basement faulting (Fig. 2.17C: Tankard et al., 1982, p. 411) and is overlain by the shale, sandstone and local conglomerate of the Kirkwood Formation, which contains Lower Cretaceous plant remains (McLachlan and McMillan, 1976; corresponding with the Wood Beds which W. G. Atherstone recognized in the middle of the last century). Overlying these are the estuarine siltstone and shale with very subordinate limestone, overlain by the shallow-marine Sundays River Formation. The marine part of the Kirkwood and the Sundays River Formations have an age range of late Valanginian to Hauterivian, although some Kirkwood faunas have latest Jurassic affinities (McLachlan and McMillan, 1976). These formations also interfinger landward, indicating generally a rift-perforated piedmont environment grading into a shelf southward.

The rifts that form along large strike-slip segments may also be sphenochasms related to drag-related tearing of one of the future continental margins. The Cenozoic rifts of northwestern China around the

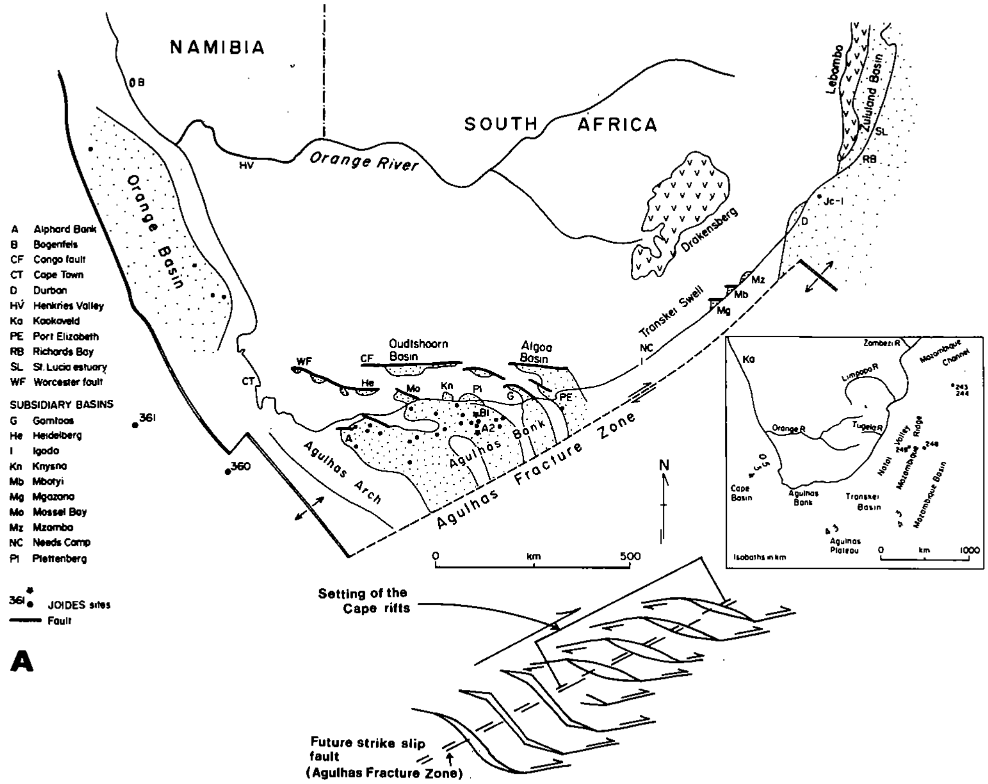
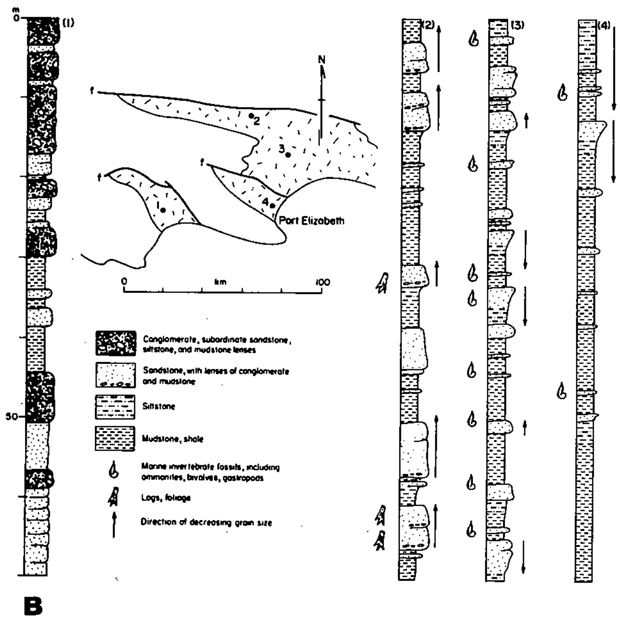
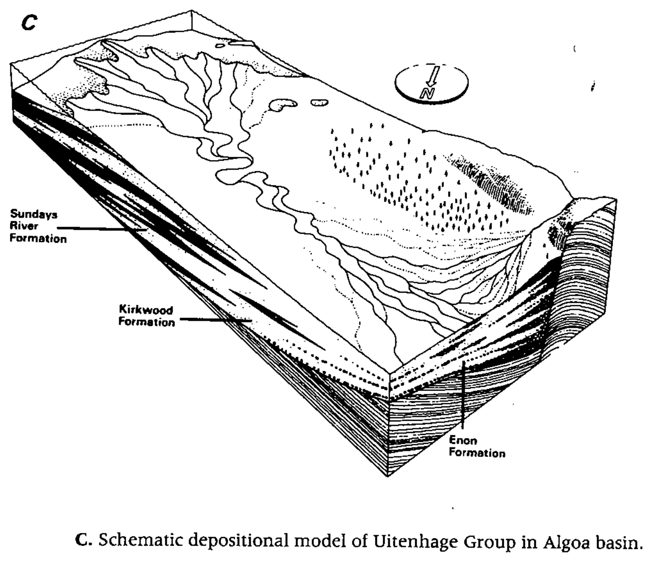


Figure 2.17 Cape rifts as examples of aulacogens that formed through strike-slip-related secondary extension (path d2-k3-k33-g1 in Fig. 2.5). (Reproduced with permission from Tankard et al., 1982, Figs. 12-2, 12-4, and 12-5.)

A. Map showing distribution of Mesozoic aulacogens in South Africa and their spatial relation to strike-slip margin formed by Agulhas fracture zone. Inset shows my interpretation of mechanism of formation of South African aulacogens as a combination of rotated tension gashes joined at their tips to Riedel shears (Modified from Wilson, 1960, Fig. 3B).



B. Sedimentary rock types deposited in Cape rifts, as exemplified by four sections observed in Algoa and Gamtoos basins (for location of the basins see Fig. 2.17A). (1) Enon Formation in Gamtoos basin; (2) and (3) Kirkwood Formation, and (4) Sundays River Formation in Algoa basin.



C. Schematic depositional model of Uitenhage Group in Algoa basin.

Ordos block may be a modern example of sphenochasms forming along the Qin Ling strike-slip fault systems that do not extend beyond the Tan-Lu fault (Fig. 2.18; Ma et al., 1982; Burchfiel and Royden, 1991).

6. Continental-rotation hypothesis (path (d1 or d2)–k2–(k21 or k22)–(g2 or g3) in Figure 2.5: products commonly aegeotype).

Yet another way in which an aulacogen may form is shown schematically in Figs. 2.11J–K. If a continental piece were rifted and rotated for a limited amount from the main continent, perhaps similar to Iberia's rotating away fom the rest of Europe to open the Bay of Biscay (e.g., Ries, 1978), then the resulting rift scar (Fig. 2.11K) may be left as an aulacogen. For example, Burke and Dewey (1974) suggested that the Benue Trough and the associated central African rift system (cf., Burke, 1976; Fairhead, 1988) may have formed in response to an aborted attempt by northwestern Africa, in the latest Jurassic - Early Cretaceous, to rift and rotate away from the rest of the continent. This attempt (cf., also Pindell and Dewey, 1982) created a very complex rift net extending from the Benue in Nigeria to Chad in the north and to the Sudan in the east.

The sedimentological characteristics of the aulacogens resulting from this mechanism depend on whether the original rifting and rotation resulted from active (*d1*) or passive (*d2*) modes of rift formation and are similar to those of the corresponding aulacogen types.

7. The hypothesis of collision-related rifts that become aulacogens at the same time (path d2–k4–k42–(k421–k422–k423)–(g1–g3–g4) in Fig. 2.5: products may be of both afro- and aegeotype).

Şengör (1987a) suggested that a continent may break up upon collision with another continent-creating collision-related rifts. If any of these rifts extends as far as the uncollided margins of the continent in question, they will be "late aulacogens" with respect to that margin. One possible example that comes to mind is the Paleozoic Amazon rift that may have formed upon a Pan-African collision at its eastern end, along the São Louis craton suture in the latest Precambrian (with cooling ages well into the Ordovician: cf., Burke et al., 1978), and faced ocean at its western end (cf., Şengör and Burke, 1991). The Donets aulacogen in the East European Platform (one of Bogdanov's, 1962, "late aulacogens") may have originated as an impactogen resulting from an early Paleozoic collision along the southern part of the Tornquist-Teisseyre lineament (cf., Blundell et al., 1991).

Naturally, two or more of the mechanisms reviewed above may operate at once in the formation of one or a group of aulacogens. For example, I have already mentioned under hypothesis 6 that the structural and sedimentological character of resulting aulacogens will be dependent upon whether the rotation commenced as a result of active or passive rifting. Of the suggested hypotheses, only the membrane-stress-related rifting may be inapplicable to the formation of aulacogens.

Post-Rifting Evolution of Aulacogens

Although somewhat dependent on their initial mode of formation, the subsequent geological evolution of all types of aulacogens ideally follows similar paths. Figure 2.19 is an idealized schematic block diagram showing the essential elements of a mature aulacogen. Subsequent to formation of the ocean, into which the aulacogen opens, the continental margin carrying the aulacogen subsides and shelf sediments cover the aulacogen, at least its more "distal" parts (e.g., the Reelfoot rift: Fig. 2.3A; the New York Bight rift: Figs. 2.15C and D). The aulacogen at this stage may localize continental drainage and nucleate large deltas at its mouth, as has happened to the Mississippi Embayment aulacogen (Fig. 2.2: Cenozoic delta progradation: McGookey, 1975; see also Moore et al., 1979) and the Benue Trough (Fig. 2.2: Niger delta: Burke, 1972) (see Figs. 2.11D, G, and I; for the localization of the world's major deltas by aulacogens, see also Audley-Charles et al., 1977). Such large, pre-orogenic deltas found in the stratigraphy of an orogenic belt against a rift striking at a high angle to it may help identify the rift as an aulacogen, as opposed to an impactogen (e.g., in the Vilyuy aulacogen: Parfenov, 1984, esp. p. 38ff and Fig. 13).

Burke (1975) suggested that, because ocean opening is a diachronous event, seawater may spill from the more advanced sectors into regions still at the rift stage and form very thick salt deposits (e.g.,

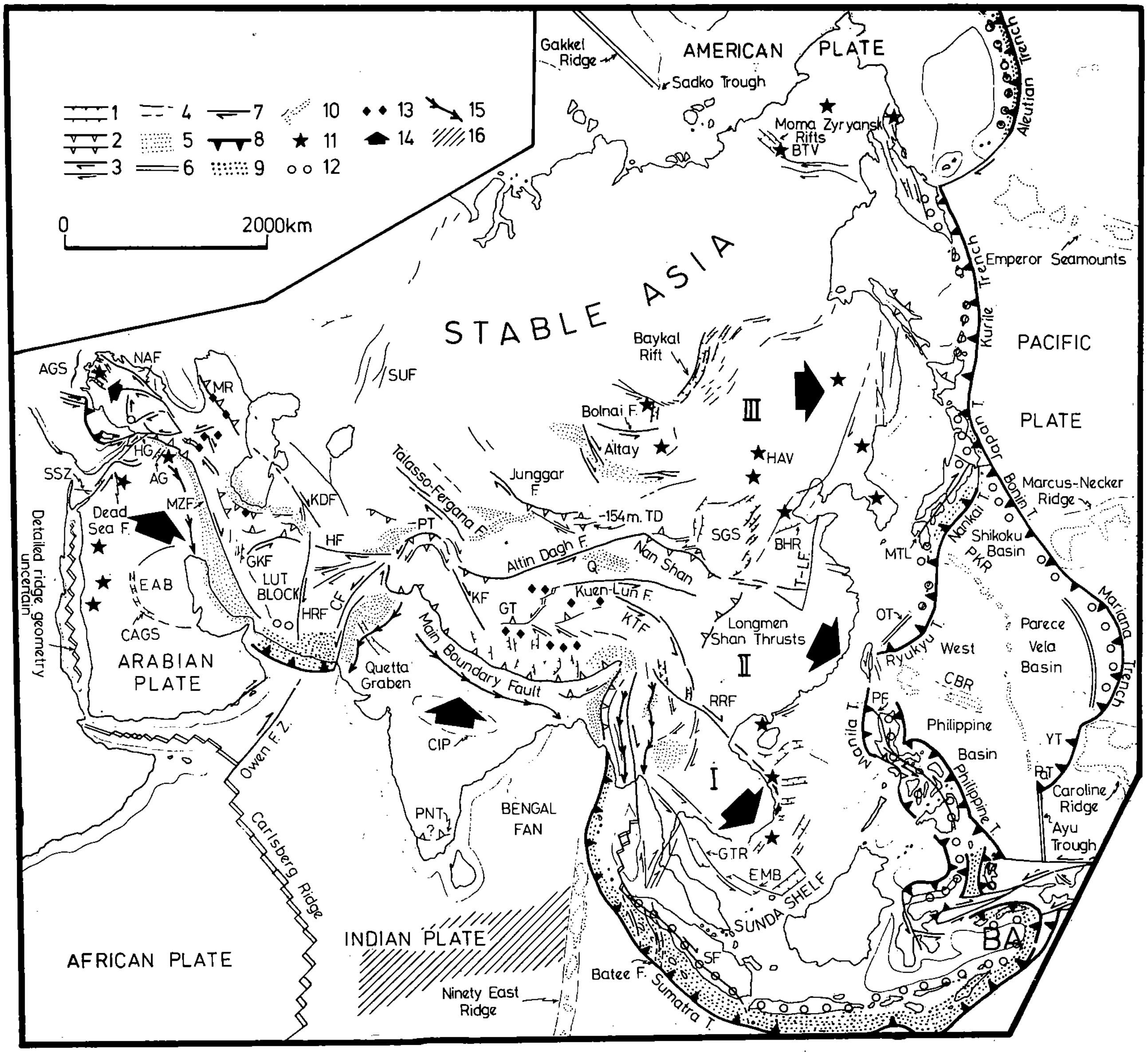


Figure 2.18 Neotectonic map of Asia, showing peri-Ordos sphenochasms ("Shansi Graben System": SGS) that were opened by the locally inhibited movement of Nan Shan block with respect to northern Asia, Tibetan rift cluster, Baykal impactogen, and Moma Zoryansk rifts. (Reproduced from Şengör, 1987d, Fig. 10.)

Key to legend: 1, first- and second-order normal fault; 2, first- and second-order thrust fault; 3, first- and second-order strike-slip fault; 4, unspecified and/or suspected fault; 5, area of folding (commonly superficial); 6, sea-floor spreading center; 7, major transform fault; 8, active subduction zone; 9, active subduction-accretion complex; 10, aseismic submarine ridge; 11, intraplate volcanism; 12, subduction volcanism; 13, "Tibet-type" collisional volcanism; 14, approximate direction of movement relative to stable Asia of various continental fragments in Asia; 15, major course of present drainage.

Key to lettering: AG, Akcakale rift; AGS, Aegean rift system; BA, Banda Sea; BHR, Bo Hai rift; BTV, Balagan-Tas volcano; CAGS, Central Arabian graben system; CBR, Central Basin Ridge; CF, Chaman fault; CIP, Central Indian plateau; EAB, East Arabian Block; EMB, East Malaya basin; GKV, Great Kavir fault; GT, Gerze thrust; GTR, Gulf of Thailand rift; HAV, Hsing An fissure volcanoes; HF, Herat fault; HG, Hatay rift; HRF, Harirud fault; KDF, Kopet Dag fault; KF, Karakorum fault; KTF, Kang Ting fault; MR, Main Range of the Greater Caucasus; MTL, Median Tectonic Line; MZF, Main Zagros fault; NAF, North Anatolian fault; OT, Okinawa Trough; PaT, Palau trench; PF, Philippine fault; PKR, Palau-Kyushyu ridge; PNT, Palni-Nilgiri Hills thrust; PT, Pamir thrust; Q, Qaidam basin; RRF, Red River fault; SGS, Shansi graben system; SF, Semangko fault; SSZ, Sinai shear zone; SUF, South Ural seismic faults; TD, Turfan depression; T-LF, Tan-Lu fault; YT, Yap Trench.

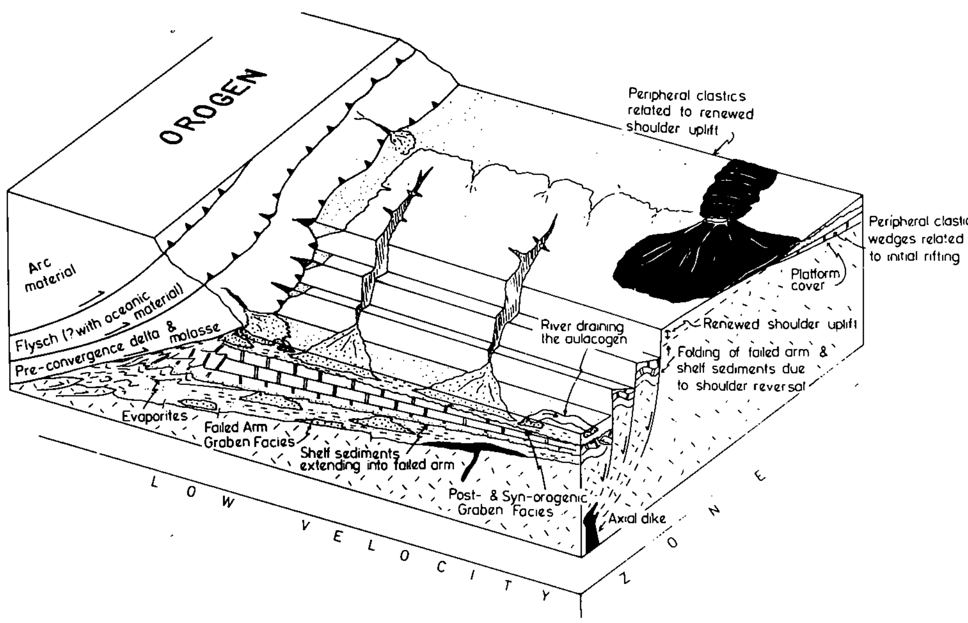


Figure 2.19 Idealized schematic diagram illustrating anatomy of an aulacogen. Subaerial erosion is ignored to emphasize structure and reactivation is assumed to revitalize original faults to keep diagram simple. See text for discussion. (Reproduced with permission from Şengör, 1987a, Fig. 4.)

in the "Mucuri Low" between the Cumuruxatiba and the São Mateus-Santa Barbara highs in the Espirito Santo basin in the Brazilian offshore, Fig. 2.2: Asmus and Ponte, 1973, Fig. 9). One of the most important roles of such evaporite horizons seems to be the trapping of oil and gas generated and preserved within the graben facies (e.g., in the Dnyepr-Donetz aulacogen, Fig. 2.2: Mirtsching, 1964, p. 127).

Whether or not doming predates rifting, rift formation is almost always followed by the uplift of rift shoulders owing to isostasy and elastic bending of the faulted plate and to extension-related pressure drop and consequent adiabatic partial melting and volumetric expansion in the underlying mantle. After the rift becomes inactive, uplift caused by the second mechanism commonly reverses and leads to subsidence, much like the young Atlantic-type rifted continental margins. This reversal may induce a very feeble axis-perpendicular compression of the rift contents and result in gentle axis-parallel folding (Figs. 2.19 and 2.20; see Borissjak, 1903; for many other examples see Milanovsky, 1981, esp. Fig. 24; Khain and Michailov, 1985, Fig. 52). However, a detailed consideration of the history of the Eurasian aulacogens has shown that axis-parallel folding of aulacogen strata is usually a result of collisional orogeny near an aulacogen with an orogenic front parallel with the aulacogen axis (e.g., the Donetz folding phases in the late Paleozoic, middle Jurassic, and early Cenozoic correlate with Hercynian, Cimmeride and Alpide collisions to the south of the aulacogen; for this and other Eurasian examples, see Şengör, 1984; see also Siedlecka, 1975, for the Timan aulacogen/Ural orogen relationship).

When orogeny begins at a continental margin bearing an aulacogen, either owing to initiation of subduction, or the onset of continental collision, the aulacogen enters a new phase of its evolution. In Fig. 2.19 it is assumed, for graphic simplicity, that this reactivation is in the form of renewed rifting. In this particular case, this phase of the aulacogen is not dissimilar to the earlier phase of an impactogen (compare Fig. 2.21Ca with 2.21J: *path k4–k42–k422–g1* in Figure 2.5). Normal faults associated with this renewed rifting were drawn to be precisely the same as those that had initially created the aulacogen. Of course, this is an unlikely situation (but see the discussion in Etheridge, 1986), and is used here only to illustrate the principles involved.

The renewed-rifting phase has the same characteristics as the previous one, such as alkaline to peralkaline (or, if the extension is considerable, even tholeiitic) magmatism, new graben-facies development,

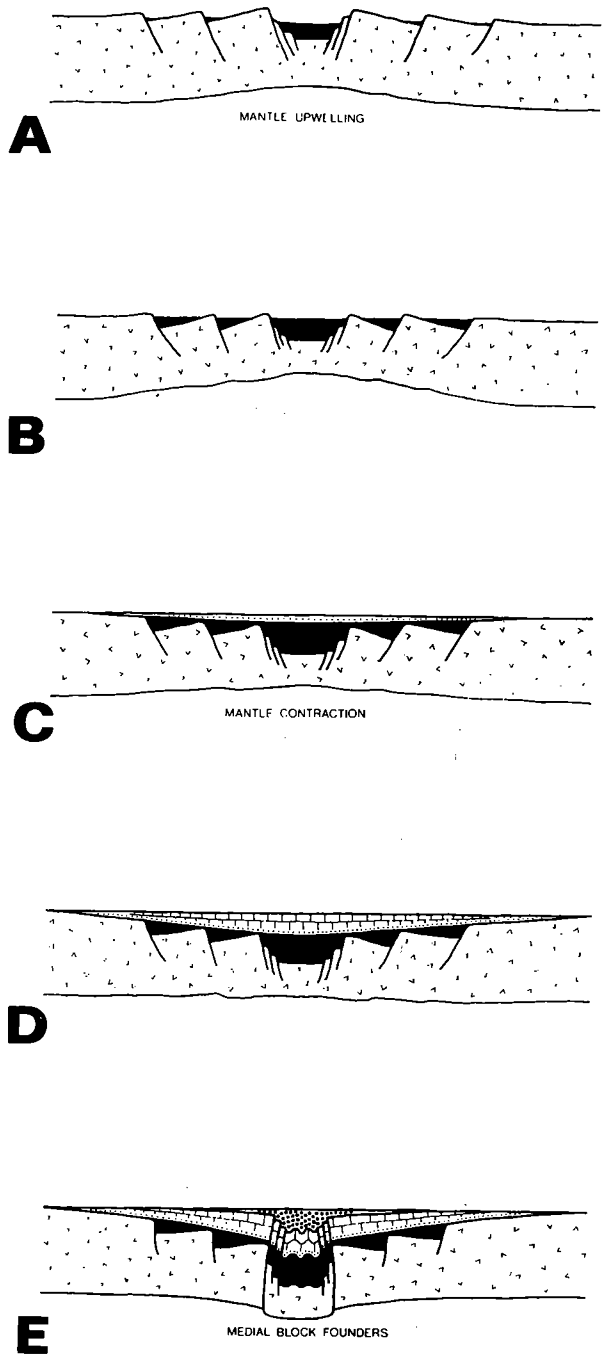


Figure 2.20 Cross-sectional evolution of Athapuscow aulacogen, showing shoulder subsidence-related shortening of rift contents. Compare this with Figs. 7 and 8 in Hoffman, 1973, showing the actual structural geometry in the aulacogen now. (Reproduced with permission from Hoffman et al., 1974, Fig. 10.)

and renewed shoulder uplift. An important difference between the sedimentary regimes of the initial rifting of the aulacogen and its "orogenic" extensional reactivation commonly is that the new mountain range provides an independent sediment source and induces a sediment transport direction along the axis of the aulacogen, opposite to the major axial sediment transport direction prevalent during the initial formation of the aulacogen. In the case of the Athapuscow aulacogen in Canada (Fig. 2.2; Hoffman, 1973) and the Southern Oklahoma aulacogen (Fig. 2.2; Ham and Wilson, 1967), coarse clastic sediments eroded from the rising collisional orogen were deposited in a trough larger than the original aulacogen. A similar situation exists in the Vilyuy aulacogen that was reactivated during the Verkhoyansk collision (Fig. 2.2; see Meyerhoff, 1982).

Reactivation of an aulacogen may happen by compressional deformation, if the aulacogen strikes at low to moderate angle to the direction of convergence of the associated orogen (Fig. 2.21Cc). The Timan mountains in northwestern Russia, a compressed aulacogen related to the terminal Ural collision, is an example of such compressional reactivation (Siedlecka, 1975). Strike-slip faulting parallel or subparallel with the axis of an aulacogen may also occur during reactivation of the structure under the influence of a nearby orogeny, again depending on the angle between the axis of the aulacogen and the direction of convergence (Fig. 2.21Cb). If the angle is fairly high, this strike-slip event may accompany some extension and be expressed as a transtensional reactivation. If the angle is moderate, it may be transpressional. In the Southern Oklahoma aulacogen, Wickham et al. (1975) recognized such large-scale strike-slip faulting, parallel with the aulacogen axis, related to Ouachita collisional orogenesis.

Finally, one of the more important characteristics of aulacogens that was first noted by Borissjak (1903) and Shatsky (1955; see Schatski, 1961, p. 175–196) is their *repeated reactivation* (see also Milanovsky, 1981, esp. Fig. 24). This characteristic has long been used as an argument for their "deep roots" reaching into the mantle and thus precluding continental drift, especially by certain schools of Soviet geologists (e.g.,

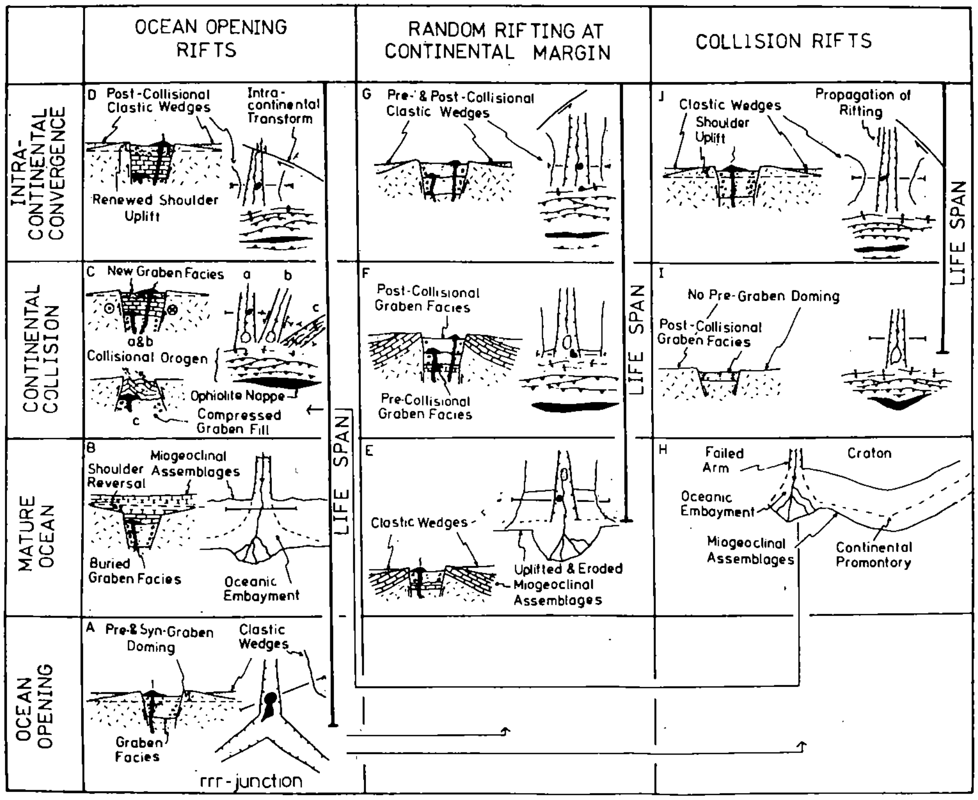


Figure 2.21 A schematic illustration of events during origin and evolution of an aulacogen (**A-D**), a random rift at high angle to a continental margin (**E-G**), and an impactogen. Cross sections show expected differences in stratigraphic evolution of three kinds of "high angle" rifts. Note that in **C**, three differently trending aulacogens, with respect to their related continental margin, are illustrated to show varied response they display to collision (ranging from re-rifting at **Ca** to compression across axis at **Cc**; reproduced with permission from Şengör et al., 1978, Fig. 1).

Beloussov, 1980, esp. p. 297ff). But repeated reactivations of aulacogens is most conveniently explained in terms of Wilson's (1968) model of repeated ocean closure, opening and reclosure along roughly the same lines (not necessarily along the same *locations*, an incorrect view that some "terrane analysts" have tried recently to ascribe to Wilson! See Şengör, 1990c, p. 21–22). Repeated ocean opening and closure (i.e., repetitive Wilson cycles) cause marginal structures such as aulacogens to nucleate new rifts. This situation also introduces the added complexity of confusing reactivated aulacogens with collision-induced rifts. I discuss field criteria to distinguish aulacogens from collision-related rifts below, after I review the geology of the latter.

Repeated compressional reactivation of aulacogens also frequently occurs if orogenic belts of different ages evolve parallel with one another and with the trend of the aulacogen, along the margin of the continent on which the aulacogen is located (e.g., Donetz aulacogen described above). Thus, repeated compressional reactivation of aulacogens is also caused by Wilson-cycle-related events along other margins of the continents.

Aulacogens are important structures because they preserve the valuable pre- and syn-rifting stratigraphic record of now vanished oceans. Such early records are usually obliterated during orogenies that eventually consume the oceans. Aulacogens are also economically important, as they commonly contain rich economic reserves, such as hydrocarbons and rift-related minerals, as well as rich evaporite deposits.

Continental-Collision-Related Rifts

Brief History of the Concept

That orogeny leads to rifting at high angle to the orogenic trend is a twentieth-century idea (Fig. 2.22). Weber (1921, 1923, 1927) first pointed out that joint sets perpendicular both to axial planes of folds and to fold axes in orogenic belts are commonly seen to propagate into forelands, where they bound extensional structures. He concluded from this that compression caused by orogeny also affects the cratons, forming forelands, and leads to their extension at high angle to the compression direction.

In the early days of plate tectonics, rifting was viewed almost entirely within the framework of continental fragmentation and was associated with divergent plate boundaries, parallel with Wegener's initial interpretation of rift valleys in terms of continental breakup. Only in the mid-seventies did it become fashionable *again* to consider rifting also as associated with orogeny, especially continental collision, and here it was once more the German school that took the lead (e.g., Illies, 1972, 1974a,b, 1975a,b; Illies and Greiner, 1978).

In 1975, Molnar and Tapponnier interpreted the Lake Baikal rift as a product of collision-related extension in terms of a plastic-rigid indentation model for Asia (Molnar and Tapponnier, 1975; also

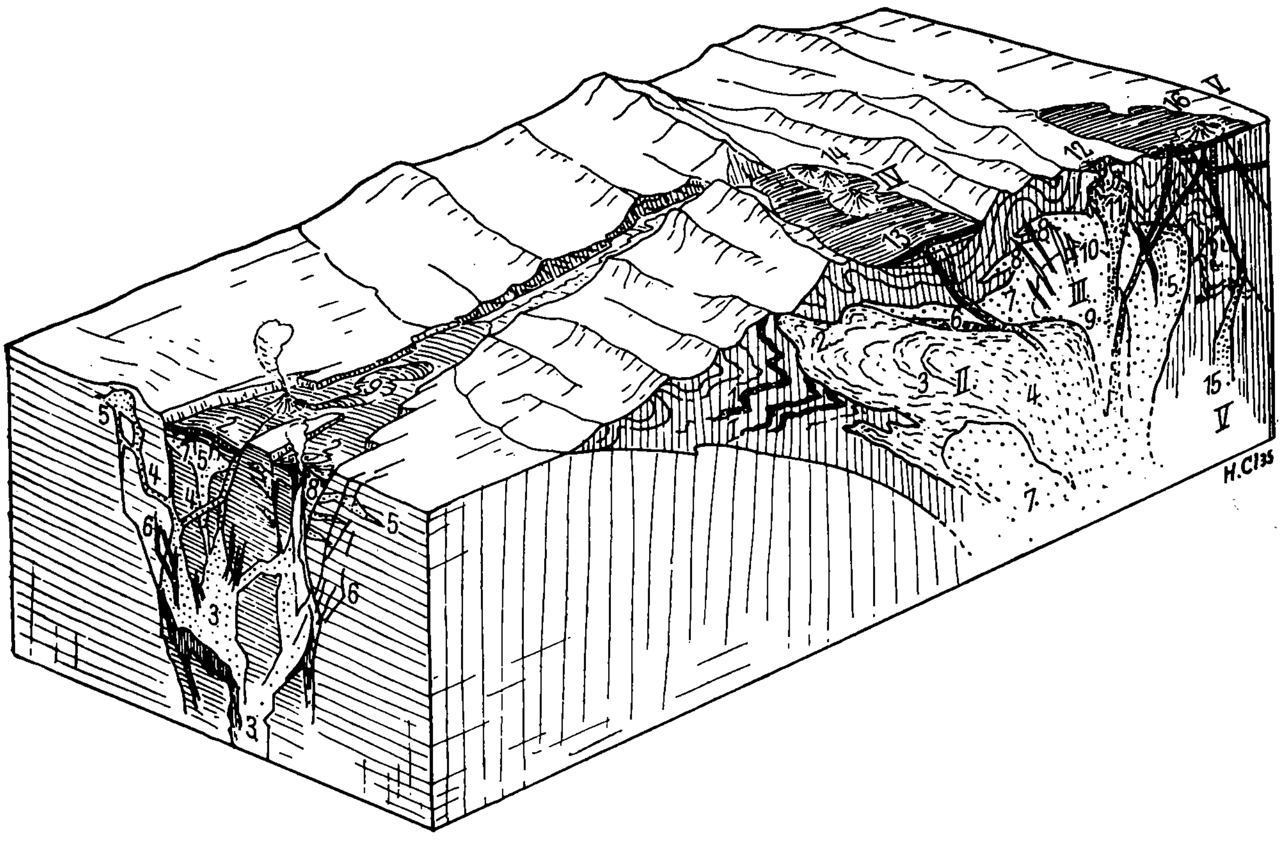


Figure 2.22 A collision-related rift, as illustrated by Cloos (1936, Fig. 120, reproduced with permission) to illustrate related magmatism (Cloos' interpretation of the origin of the rift was different, as explained in text, but the way he drew the orogen and the rift speaks for itself!). Key to numbers (all italicized annotations are mine): Left: Alkalic rocks in regions of extension: 1. Mafic parent at depth; 2. Mafic parent at surface; 3. Intermediate and felsic melts in branched plutons; 4, 5. Intermediate and felsic melts in dikes and subvolcanic plutons; 6,7. Late melts, mainly mafic, at depth and in shallow dikes; 8,9. Late melts, mainly mafic, in volcanoes and lava flows. Right: Alkalic rocks in folded mountains: I. 1. Mafic and ultramafic early melts at bottom of geosyncline (ophiolites, diabases, etc.) (*we now know that ophiolites are floor remnants of vanished oceans and they include diabase here separated by Cloos!*); II. 5. Intermediate and felsic plutons of first, concordant, synorogenic main magmatism (*i.e., calc-alkalic arc-related magmatism in our present terminology*); III. 6. Mafic early magmas; 7,8,9. Intermediate and felsic plutons of second, discordant, late-orogenic main magmatism (*i.e., calc-alkalic late arc and collision-related magmatism in our present terminology*); 10. Late dikes; 11,12. Late-orogenic subvolcanic plutons and volcanoes (*post-collisional i.e., Tibetan, and/or pre-collisional i.e., Altiplano-type plateau magmatism*); IV. 13,14. Post-orogenic magmatism in intramontane basins (*e.g., in Eastern Turkey: see Pearce et al., 1990; Yılmaz, 1990*); V. 15, 16 Postorogenic magmatism in the hinterland, partly becoming alkalic (*"orogenic collapse" magmatism: cf., Lorenz and Nicholls, 1976*).

see Tapponnier and Molnar, 1986, esp. Fig. 4a). Şengör (1976a,b) proposed that continental collision commonly leads to rifting at high angles to the collision front and Şengör et al. (1978) termed such rifts *impactogens*, although not much attention originally was paid to the existence of various classes of collision-induced rifts with different genetic significance; thus, in many subsequent publications, all collision rifts were loosely referred to as "impactogens" (e.g., Burke, 1980, p. 45), until Şengör (1987b, p. 336) made a distinction between impactogens, which are "only those high angle rifts that result from extensional strain induced *directly* by continental collision without an intermediary process", and other collision rifts. This definition, thus took out of the impactogen concept those rifts that form in relation to other structures related to collision, such as strike-slip-related rift basins (e.g., the Akçakale Graben in SE Turkey: Tardu et al., 1987, esp. Figs. 1, 5a and 19).

In the eighties, the concept of collisional rifting grew in popularity and was not only applied to places where it originated, but also elsewhere (for reviews see Hancock and Bevan, 1987 and Şengör, 1987b), even including cases of ocean opening as a result of collisional rifting (e.g., Sawkins and Burke, 1980). In the following subsections, I review the tectonic and sedimentary relationships in the three main types of collision-induced rifts, namely impactogens (k421), rifts that form along intracontinental convergence belts (k422) and pack-ice-type rifts (k423).

Impactogens

Origin of Impactogens

Impactogens are thought to be produced by secondary tension set up at high angles to a zone of compression, similar to the situation in a "Brazilian" tensional strength test performed by loading a cylinder across a diameter between flat surfaces applying concentrated loads (Jaeger and Cook, 1971, p. 160).

Şengör (1976a,b) suggested that this "parting" mechanism (cf., Beloussov, 1962, p. 542, esp. Fig. 233) was responsible for the formation of impactogens. That the bounding faults of large impactogenal rifts are not vertical tension fissures, but normal faults, is a result of gravitational force that does not significantly affect fractures forming in cylinders in experiments.

The model suggested by Molnar and Tapponnier (1975) and Tapponnier and Molnar (1976) is similar in principle, but assumes plastic, instead of elastic, behavior of the failing medium. If this medium is of finite extent (as continents such as Asia are) then a tensile state of stress will arise in the region adjacent to the boundary of the plastic medium opposite an indenting die (in our case, a colliding continent such as India).

Both Şengör's (1976a,b) and Tapponnier and Molnar's (1976) suggested mechanisms for impactogen formation may be valid, depending on the thermal state of the strained lithosphere and the rate of strain. Indeed, Şengör's (1976a,b) model was based largely on the Cenozoic tectonic history of the Alpine foreland in central Europe, where an initially cold lithosphere (last alpinotype orogeny was of late Paleozoic age) had been affected by a fairly rapid Alpine collision (see Trümpy, 1972, with the reservations expressed in Şengör 1991a). By contrast, Tapponnier and Molnar's (1976) model was developed on the basis of the Cenozoic tectonics of the Himalayan hinterland in central Asia, which consisted not only largely of Paleozoic/Mesozoic soft subduction/accretion material (cf., Şengör and Okuroğulları, 1991; Şengör, 1992), but also of blocks that had been affected by *late Mesozoic* alpinotype orogenies all the way from the southern tip of the Baikal rift to the Himalayan suture zone (see Şengör, 1987d; Şengör and Okuřogullari, 1991). This substrate is thus expected to behave more in a plastic fashion than the cold Alpine foreland, and perhaps, to induce a lower strain rate by making the straining area wide, despite the much higher convergence velocity of India with respect to Asia than was the case in the Alps.

Figure 2.23 summarizes in a schematic fashion ways in which impactogens are thought to originate in the framework of Şengör's model of elastic parting and some complications. Figure 2.23A displays two continents, *a* and *b*, that are converging at the expense of an intervening ocean that is being subducted along a trench located at the foot of the continental margin of continent *b*. In Fig. 2.23B, the promontory of continent *a* has collided with continent *b*. At this stage, the

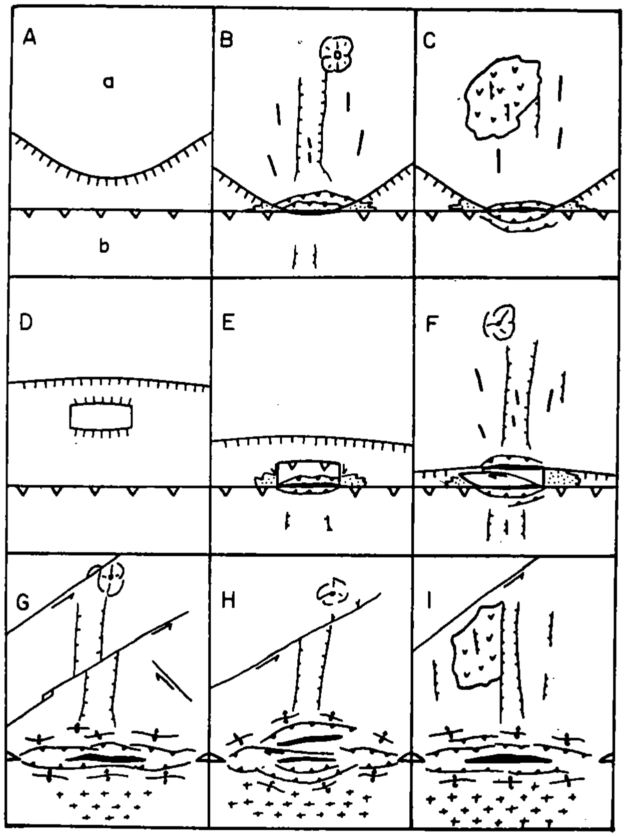


Figure 2.23 Sketch maps showing generation of Upper Rhine-type impactogens and some complications. The sequential maps are as follows: **A,B,G; A,C,I; D,E,F,H**.

Key to ornaments:

Lines with long hachures: Atlantic-type continental margin, hachures on the ocean side; lines with open triangles: Andean-type continental margins, triangles on overriding plates; lines with short hachures: normal faults with hachures on downthrown side; short, thick, black lines: mafic dikes; lines with black triangles: thrust faults with triangles on upper plate; dotted areas: flysch wedges; v's: flood basalts; lines with half-barbed arrows: strike-slip faults; black regions: ophiolitic sutures; lines with double-headed arrows across them: anticlines; plus signs: regions of Tibetan-type basement reactivation. (Reproduced with permission from Şengör, 1987b, Fig. 1)

collision is only effective at the *point* where the promontory has touched the continental margin of *b*. Such *point collisions* (in reality *line collisions* when one considers the thickness of the colliding continents) are analogous to the line loading of cylinders along their diameters in the Brazilian test discussed above, and similarly, produce tensional stresses perpendicular to the direction of convergence. If the collision is head-on, then rifting will occur perpendicular to the collision front, and thus generate a high-angle rift (cf., Şengör et al., 1978). Figure 2.23B exhibits a situation in which neither of the collided margins has been thermally (by magmatism) or mechanically (by the growth of large accretionary complexes) weakened by long-lived subduction activity. This may be the case where the ocean that disappeared during continental collision had a small cross-convergence width (cf., Şengör, 1991b), as was likely the case in the Alps (<1000 km: Şengör et al., 1984; Şengör, 1991b). The collision of two such "strong" continental margins would result in rapid communication of the collision-induced stresses to the fore- and hinterlands and give rise to their deformation by convergence-perpendicular extension.

The structural expression of such deformation may be extension joints, mafic dikes and gravity faults, probably in that order of occurrence (Fig. 2.23B). In the case of the Upper Rhine rift, Bergerat (1983, 1987) and Villemin and Bergerat (1985, 1987) were able to document that the earliest structural event in the formation of the future Upper Rhine rift was north-south shortening, as expressed by NNE-SSW-striking conjugate strike-slip faults, NNE-SSW-striking tension gashes filled mainly with crystalline calcite, and E-W-striking horizontal stylolites. Although Bergerat and Villemin were unable to date this event with any precision, they did establish, by cross-cutting relationships, that these structures were the earliest in the history of the Upper Rhine rift. However, they noted that some of their tension gashes displayed a field continuum with some dikes (Villemin and Bergerat, 1985, p. 66). Illies (1974a, esp. Fig. 3) had earlier pointed out that the earliest harbinger of rifting along the present Upper Rhine structure was olivine-nephelinite magmatism that created dikes between 65 and 55Ma, whose broad locations and strikes are identical with those of the tension gashes mapped by Villemin and Bergerat. Data reported by Illies (1974a) date the earliest phase of the Rhine rift taphrogeny as Paleocene-Eocene. On the basis of regional considerations, Bergerat and Villemin also placed this phase in the Eocene.

Sedimentation that was coeval with this phase of tectonism was recorded by the 100m-thick fluviatile-limnic clastic rocks and rarer limestone and dolostone

(Doebl, 1970). The point emphasized by Bergerat (1983, 1987) and Villemin and Bergerat (1985, 1987), that pronounced normal faulting did not take place at this time, is indicated by the modest thickness of these early sediments and by the fact that they are distributed in an area larger than the future rift trough (Illies, 1977, p. 335). Illies (1977) pointed out, however, that even the earliest of the Rhine rift sediments show evidence for some normal dip-slip faulting.

The oldest sedimentary rocks of the Rhine rift, of Lutetian-Bartonian age on the basis of mollusc and vertebrate fossils, are dominantly fluviatile conglomerate at the base (clast composition being strongly controlled by the underlying rocks) and passing upward into claystone and sandstone with rarer marl; locally, there are limnic limestone and dolostone (Doebl, 1970). Freshwater limnic sediments of Lutetian age, distributed in an area not exactly coincident with the later rift trough, were laid down next, and contain freshwater limestone, marl, some dolomitic marl, rare sandy limestone, and lignite, indicating the presence of palludal environments (and thus limited topographic differentiation). The thickness of all these rocks nowhere exceeds 50 m and supports the inference that the rate of normal faulting was modest, in agreement with Bergerat and Villemin's structural observations.

By Priabonian time (late Eocene), rates of normal dip-slip faulting and related subsidence accelerated. The lower parts of these sequences consist of marl, marly limestone, anhydrite, and salt. In their upper parts, there is dolomitic marl (Sittler, 1965; Doebl, 1970). Toward the master faults of the rift trough, the rocks are conglomeratic. The thickness of the Priabonian sequence is up to about 900m to the southwest of Freiburg i.Br., about 500m near Karlsruhe, and close to zero near Mannheim (Fig. 2.24). Upper Eocene strata also indicate the first marine incursion into the rift trough that laid down the thick (about 300m) salt deposits.

Late Eocene subsidence correlates temporally with strong roughly east-west extension, indicated by numerous normal dip-slip faults and normal dip-slip movement on former strike-slip faults accompanied by continued opening of tension gashes (Bergerat, 1983, 1987; Villemin and Bergerat, 1985, 1987). Villemin and Bergerat (1985, p. 68) pointed out that, among the normal dip-slip faults, easterly dips predominate. This observation is in excellent accord with the more pronounced subsidence at this time along the western margin of the rift trough, interpreted by Illies (1970, esp. Fig. 5) as asymmetric rifting by the formation of the (east-dipping) western master fault system first.

If the active continental margin in the collision system is a well developed Andean-type arc (e.g., Fig. 2.23C), then subduction-related magmatism is likely to have weakened it before the collision (cf., Armstrong, 1974; Burchfiel and Davis, 1975; Isacks, 1988; Şengör, 1990b; Dewey and Lamb, 1991, esp. Fig. 4). This leads, during the collision, to absorption of much of the convergent strain by the shortening of this "weak" margin. As a result, at least during the early stages of the continental collision, the foreland structures described in the preceding paragraphs may be only weakly developed. They may, for example, be confined only to a few gravity faults and dikes feeding flood basalts (Fig. 2.23C), as exemplified by the southeast Turkish foreland on the Arabian plate. Here, the soft Andean-type margin of Eurasia (cf., Yılmaz, 1985, 1990) took up much of the convergent strain by shortening and complementary thickening (cf., Şengör and Kidd, 1979; Dewey et al., 1986). The only foreland structures that developed in an impactogenal fashion are the N-S-striking fissures that fed the large Plio-Pleistocene basaltic rocks of the Karacadağ shield volcano (Şengör et al., 1978; Hancock and Bevan, 1987, Fig. 1b). Similarly, no major impactogens are known from the Indian subcontinent that collided with a rather "soft" Eurasian margin (Harrison et al., 1992). In the case of the latest Carboniferous - earliest Permian Oslo rift (Fig. 2.2), the earliest products of a much delayed impactogenal(?) rifting were the feeders of the alkalic basalt and rhomb-porphyry eruptions (Ramberg, 1976, p. 169).

Figures 2.23D–F show how a microcontinent between two larger converging continents in a collisional system may influence impactogen generation. Although in Fig. 2.23D there is no continental promontory to cause "point collisions", Figs. 2.23E and F show that the betwixt microcontinent functions as just such a promontory. The difference

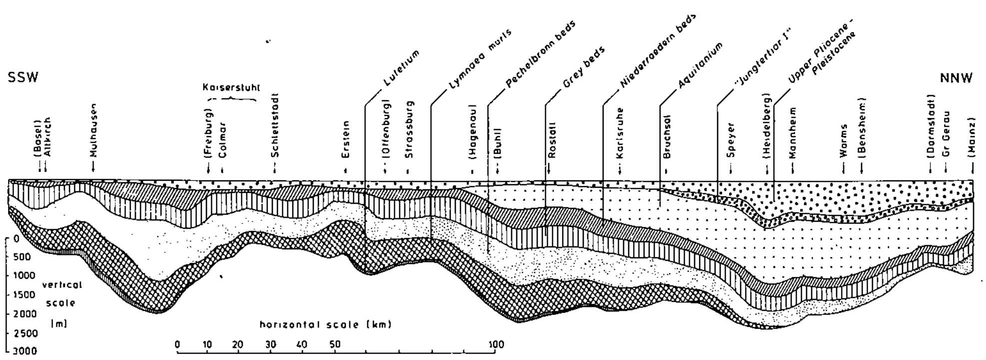


Figure 2.24 Thickness distribution of Tertiary sedimentary units in the main trough of the Upper Rhine rift displayed along a longitudinal section. As this figure illustrates, rift-related subsidence commenced in the south during Lutetian (medial Eocene). The center of maximum subsidence shifted northward during later evolution of the rift. Accelerated rift-floor subsidence is seen for the late Eocene - early Oligocene interval (Lymnea marls and Pechelbronn Beds), for Aquitanian (early Miocene), and late Pliocene-Pleistocene (for the last interval, see also Fig. 2.9B). (Reproduced with permission from Illies, 1975, Fig. 1.)

between the final picture that will result from the evolution shown in Fig. 2.23A and that in 2.23D is that in the former case the final picture contains a single suture (Fig. 2.23G), whereas in the latter, a double suture is seen (Fig. 2.23H). Such a case may have happened in the Alps, where the Briançonnais betwixt continents (cf., Trümpy, 1980, Fig. 15; Hsü and Briegel, 1991, Fig. 12.10) may have created a "promontory effect", where in reality, there was an embayment in the European continental margin (see esp. Hsü and Briegel, 1991, Fig. 12.2, frame 60 Ma).

Impactogenal rifting in the plastic-rigid indentation model of Tapponnier and Molnar (1976) also follows a similar evolutionary path as one that resulted from elastic parting, here described for the Upper Rhine rift. The Lake Baikal rift, for instance, seems to have opened in the early to medial Eocene (55 to 45 Ma), as suggested by the sudden deposition in its southern sub-basin of conglomerate, sandstone, and shale, whose cumulative thickness is 1100m (Grossheim and Khain, 1967). In the same places, Paleocene thicknesses were, at most, tens of meters. A shallow lake also formed around the present southwestern tip of the rift. In the late Eocene, similar conditions persisted, and in the Oligocene, the two distinct sub-depressions of the present rift basin, separated by the Akademichesky Ridge, became clearly recognizable. To the SE of the Akademichesky Ridge, 1000m of sandstone with some conglomerate were deposited, indicating that the southern sub-basin of the rift was the main venue of rifting in the later early Cenozoic (cf., Grossheim and Khain, 1967). Even if one followed Zonenshain and Savostin's (1981) slightly younger dating, rifting in the Lake Baikal area seems to have happened immediately after (if not synchronously with, if the later dating of the Himalayan collision at 45 Ma by Dewey et al., 1989a, is accepted) the closure of Neo-Tethys along the Indus-Yarlung suture.

It thus seems that impactogens form either synchronously with the collisions that give rise to them or that they follow after a short time, perhaps not longer than a maximum of 20 my (obtained by assuming that the Himalayan collision occurred at 55 Ma and that the Baikal rifting commenced during the Oligocene; i.e., at about 35 Ma).

The scheme of impactogen generation sketched above glosses over factors that may assist, complicate, or hinder impactogen generation. Şengör (1976b) summarized these as follows:

1. Generation of impactogens may be dependent on the rate of convergence of the two colliding continents and on the angle of collision. High rates of convergence and high angles of convergence with respect to convergent boundaries will assist impactogen generation.

2. The existence of zones of weakness in the fore- or hinterlands may nucleate new structures. Such

weak zones may prescribe, to a certain degree, their extent and orientation, and therefore, complicate the picture. Illies (1962), for example, pointed out that the orientation of the Upper Rhine rift had been influenced both by an Hercynian shear zone and some possible middle Jurassic rifting(?) (for the data see Illies, 1977, Fig. 6) only in the southern part of the future rift.

3. If crustal and/or lithospheric domes exist in the fore- or hinterlands, they may localize the formation of impactogens. These domes help impactogen formation in two independent ways. One is through the potential energy they store and thus predispose the rocks to rifting. The second is that they weaken the lithosphere either by crustal thickening or by lithospheric thinning (thus increasing the quartz+ feldspar/olivine ratio in any given cross section across them). These effects are thus cumulative. Since Cloos' (1939) classic study, all authors who have written about the Upper Rhine rift have pointed out that the Late Cretaceous swelling of the "Rhenic Shield" (Cloos, 1939) in one way or another must have helped the formation of the rift. Although I do not agree that the dome of the "Rhenic Shield" caused the rifting (as Burke and Dewey, 1973, and Illies, 1974a, following Cloos, 1939, thought), I certainly believe that its potential energy must have been a factor in localization of the rift.

4. The existence of a well developed continental-margin magmatic arc (i.e., "soft", convergent margin) on one or on both of the approaching margins may hinder the formation of impactogens by introducing a ductile zone between the two colliding continents, which may take up the impact of the collision (i.e., by reducing the effective rate of convergence of the brittle cratons). In other words, such a ductile zone reduces the "impact" of the collision on the brittle fore- and hinterlands.

Evolution of Impactogens

Once extension begins along the future site of an impactogen following a continental collision, rift nucleation succeeds the initial formation of individual extensional structures (Figs. 2.23B and F). Such rifts commonly form first near the collision site and propagate later away from the collisional orogen. As a consequence, the amount of total extension and correlatively, the magnitude of the total subsidence, is commonly greatest near the suture and decreases away from it. In the above subsection, the thickness changes encountered in the Priabonian section of the Upper Rhine rift from about 900 m in the south to 0 in the north near Mannheim illustrate the decrease in subsidence northwards away from the Alpine front. This difference in rifting intensity between the north and the south in the Upper Rhine survived into the earliest Oligocene. In the case of Lake Baikal rift, rifting commenced in the south, nearer the collision front, and then propagated northwestwards away from the collisional orogen. Even today, the deepest part of the Baikal rift is located in the southern half of the rift trough.

Volcanism within the rift trough is not ubiquitously present in all impactogens, being totally absent in some (e.g., Lake Baikal: Logatchev et al; 1983), and overabundant in others (e.g., the Oslo rift: Ramberg, 1976). In all cases however, it is a post-rifting phenomenon, or at least it commences together with faulting and tension-gash formation. Within the Upper Rhine rift, major volcanicity commenced with the activity of the Kaiserstuhl volcanic center in the south about 18 Ma (Illies, 1974a,b, 1975b)(i.e., about 27 my after the Rhine rift had originated) and, in the south, subsided to more than 2.5 km. The subvolcanic breccias of the Kaiserstuhl and the close correspondence of volcanism and the rift suggest that faulting controlled volcanism (Baranyi, 1974). In the case of the Lake Baykal rift, volcanism occurred outside the rift trough and commenced with the eruption of trachybasaltic lavas in the Miocene (Florensov, 1965)(i.e., at least 10 my after the commencement of rifting, which may be as much as 35 my after rifting, if rifting began synchronously with the Himalayan collision and if 55 Ma is taken as the date of this collision; see above). By contrast, in the case of the Oslo rift (Ramberg, 1976) and in SE Turkey (Şengör et al., 1978), impactogen generation was coeval with the onset of volcanism.

Synchronous with rifting, shoulders of rift(s) begin to rise, likely as a result of a combination of isostasy-driven elastic deformation on normal dip-slip faults (cf., Taber, 1927; Buck, 1988), and of partial melting and expansion of the underlying mantle as a consequence of pressure drop in the asthenosphere owing to rifting in the overlying lithospheric lid. This post-rifting rise of rift shoulders has been documented both from the Upper Rhine and the Lake Baikal rifts.

In the Upper Rhine rift, the Pechelbronn beds of dominantly evaporitic aspect were laid down in the early Oligocene (with the exception of the "Lower Pechelbronn Beds," whose age is now believed to be late Eocene: Doebl, 1970). At this time, the rifting reached as far north as the Hessen depression (Fig. 2.24; Lotze, 1971). The area of most rapid subsidence, however, was still in the south, as shown by the 1600 m thickness to the southwest of Freiburg and 900 m thickness near Karlsruhe of the Pechelbronn beds (Illies, 1974b). In Pechelbronn time, the so-called *Küstenkonglomerate* (i.e., "coastal conglomerates") began to be deposited along the master faults of the Upper Rhine rift. The important thing about these conglomerates is that they contain Jurassic material in the older pebbles, and older clasts with decreasing age of the conglomerates (Illies, 1967). In other words, the coastal conglomerates show an inverse relationship between the age of deposit and the age of its contained clasts. This indicates a progressive unroofing of the rift shoulders and the deposition of their erosion products at the foot of the normal dip-slip fault-bounded cliffs facing the rift trough. Both Illies (1970) and Şengör et al. (1978) interpreted this as late shoulder uplift.

The data I have at my disposal at this writing on the Baikal rift are not as detailed as they are for the Upper Rhine rift. Still, the equivalents of the "coastal conglomerates"do exist here too, especially along the master faults of the northern depression of the rift trough, and they are as young as medial Pliocene, indicating a very late rise of the rift shoulders here (Grossheim and Khain, 1967; Logatchev et al., 1978); the age of these deposits is now thought to be older, about medial Miocene (Dr. V. Kazmin, personal communication, 1991).

As the collision advances, the suture lengthens and the convergence becomes effective across a long front, as opposed to being concentrated at discrete nodes as shown in Figs. 2.23G, H, and I. One effect of ongoing convergence is to continue shortening the fore- or hinterlands, or both, in the direction of convergence and to extend them at high angles across the direction of convergence. This keeps the rifting active, but at a reduced pace than earlier, because the optimum stress conditions for rifting created by point collisions no longer apply. This is exemplified by the medial Oligocene evolution of the Upper Rhine rift. At this time, the graben subsidence levelled off, and a marine marl sequence was deposited all along the rift trough with more or less uniform thickness (Fig. 2.24; cf., Doebl, 1970). In the late Oligocene, the subsidence (and presumably the rate of extension) slowed down, and the deposits of this age are disconformably overlain, suggesting local episodes of erosion (Fig. 2.24). Villemin and Bergerat (1985) think that this slackening of rifting corresponded with a time of right-lateral strike-slip faulting along the rift trough, with little extension. Their structural studies indicate that the direction of maximum shortening at this time was NE-SW, slightly more easterly than the orientation of the main axis of the graben. This corresponds with the collision along the Alpine front having become effective all the way from the Pyrenees via the Provence chains into the Alps. In other words, the western part of the Mediterranean Alpides east of the Betic and the Riff cordilleras had sutured (cf., Cohen, 1980; Dercourt et al., 1986; Bethelsen and Şengör, 1993). This is also the time when the Bresse rift opened (Rat, 1975) and connected with the Upper Rhine rift via a transform fault system north of the Jura Mountains (Laubscher, 1970), and rifts in the Massif Central began opening under the influence of NW-SE-directed extension (e.g., Burg et al., 1982).

With the onset of the Miocene, two important changes occurred in the evolution of the Rhine rift: first, the center of subsidence shifted to the north, where today the thickest accumulations of Tertiary strata are found (Fig. 2.24; Sittler, 1969; Illies, 1974a,b); and second, major volcanism began in the rift trough in the south.

The Aquitanian section in the Upper Rhine rift consists of about 1.5 km of clastics and carbonates. However, after the Oligocene, the subsidence axis in the rift shifted from a north-northeast trend to almost exactly northwest (cf., Illies, 1974a,b; Şengör et al., 1978), which is reflected by the distribution of the thickness of the rift fill. The coeval structures mapped by Villemin and Bergerat (1985) also show that the maximum shortening axis shifted to a NW-SE orientation and that the Upper Rhine rift began shearing left-laterally along its axis (i.e., in a sense opposite to what was dominant in the late Oligocene) (Fig. 2.10). This led, in its middle sector, to shortening associated with thrusting of the rift shoulders onto the rift floor (e.g., near Baden-Baden: Illies and Greiner, 1976). Today, the rift as a whole is a broad strike-slip zone with associated second-order extensional and compressional features (Ahorner, 1970; Ahorner and Schneider, 1974; see also Chorowicz et al., 1989).

As the history of the Upper Rhine rift shows, while the suture lengthened and compression became effective across a whole front, as opposed to at discrete nodes, the fore- and hinterlands began to fail by conjugate strike-slip faulting. At this stage, the impactogen may or may not remain in an orientation favorable to further extension. In that case, it may undergo shear (as happened to the Upper Rhine rift), some parts may go on extending (as did the northern part of the Upper Rhine rift), and others may become compressed (as did the middle part of the Upper Rhine rift). At this point in the development of the collisional system, the entire fore- or hinterland in which the impactogen is located is generally deformed by numerous faults of various orientations and kinds, disrupting the entire area into numerous blocks (*Schollen*: cf., Dewey and Şengör, 1979, p. 84, footnote 1). That is why it is appropriate to consider their further evolution under the headings of "Rifts that form along intracontinental convergence belts" (k422) and "Pack-ice-type rifts" (k423) below.

Rifts That Form Along Intracontinental Convergence Belts

Both continental collision and other kinds of orogeny create broad belts of intracontinental convergence (cf., Şengör, 1990a). Those associated with continental collision are particularly large as exemplified by the Tibetan plateau behind the gigantic Himalayan range today (Dewey and Burke, 1973; Molnar and Tapponnier, 1975; Şengör and Kidd, 1979). Along collisional orogens, such belts of intracontinental convergence commonly form after the initial touch of the two continents, when the suture has achieved a certain length along the trend of the orogen. Along the Himalaya, for example, although the inital touch probably occurred sometime in the medial Eocene (about 55 Ma; but see Jaeger et al., 1989), the suture was not complete until sometime in the late Eocene, following at least 15° counterclockwise rotation of the Indian subcontinent (Klootwijk, 1981; see also Dewey et al., 1989).

Such intracontinental belts of convergence lengthen along their trend as they are shortened across it. In some orogens, this lengthening is simply expressed as joint sets or rare normal faults across the orogen (e.g., in the Alps: Weber, 1921, 1923, 1927; in the southern Appalachians, especially the "Gwinn-type" extensional lineaments: Gwinn, 1964; Kowalik and Gold, 1976; Krohn, 1976; Gold, 1980). In Tibet and in the Himalaya, along-trend lengthening of the orogen led to the formation of numerous rifts that trend at about 90° to the orogen (Figs. 2.2 and 2.18; Molnar and Tapponnier, 1978; Ni and York, 1978; Romanowicz, 1982; for the Himalaya, the classical example is the Thakkola-Mustang Graben in Nepal: Colchen et al., 1980; Fort, 1980, 1987, esp. Fig. 1; for the rest of Tibet, see Tapponnier et al., 1981, 1986; Han et al., 1984; Mercier et al., 1984; Rothery and Drury, 1984a,b; Armijo et al., 1986; Dewey et al., 1988; Kidd and Molnar, 1988).

The extensional basins in Tibet are probably both grabens and rifts. Many of them are asymmetric and there is no consistent direction of facing of the master faults of the basins. In fact, in many basins, facing of the master faults changes along trend. The basins are either tectonically isolated or they end on NW-SE- or NE-SW-striking strike-slip faults and form a kinematic whole with them in extending the plateau in an E-W direction (see esp. Han et al., 1984; Rothery and Drury, 1984a,b). Long, straight, continuous normal faults are rather the exception (e.g., along the Thakkola-Mustang basin: Fig. 2.2) than the rule. Most are bounded on one or both sides by discontinuous, commonly en-echelon faults with fault lengths on the order of 20 to 60 km (cf., Armijo et al., 1986). Most basins are short, rarely exceeding 100 km, but the Yadong-Gulu rift system appears like a mini "rift chain", with a length exceeding 400 km (Fig. 2.2).

The Tibetan rifts and grabens are mostly filled with latest Pliocene-Quaternary lacustrine and fluvial silt, sand and gravel deposited in alluvial-fan/channel systems. Some contain coal seams, pointing to the presence of former palludal environments, possibly indicating a slackening of tectonic activity (Armijo et al., 1986). In these basins, thicknesses rarely exceed one kilometer and this agrees well with observed maximum throws of about 1600 m along normal faults bounding the basins (Armijo et al., 1986). Figure 2.25 exhibits a graphic synopsis of the evolution of the sedimentary environments in the south Tibetan extensional basins, which are dominated by an interplay between normal faulting, and fluvial and glacial erosion. Armijo et al. (1986) have demonstrated that the rate of normal faulting in most of these grabens commonly has exceeded the rate of glacial erosion, leading to the formation of hanging valleys or valleys with a "wine-glassprofile" and enabling the assessment of crude rates of faulting.

One characteristic of extensional basins in the Himalaya/Tibet border area in southern Tibet is that they originally were avenues of a north-directed drainage carrying debris from the rising young Himalayan range onto the endorheic basin of Tibet during the Plio-Pleistocene. This changed later, perhaps in the late Pleistocene-Holocene, by stream capture led by the evolving Himalayan drainage, when the basins began draining to the south into the Ganges system (Colchen et al., 1980, esp. Figs. 2–4; Armijo et al., 1986). This imparted onto the sedimentary record in these basins a two-phase fluvial history, which is the opposite of the drainage reversal seen in aulacogens later reactivated by collisions.

The average amount of extension of the 1100 km-long Himalayan front has been 1% per million years during the last 2 to 2.5 my This indicates an average extension per basin of about 1.5 mm/y, which is well below the average rate of extension in most rifts in the world (which is about 1 cm/y). As a consequence, none of these basins forms rift valleys comparable in size with those of the other environments, except some of the extensional basins in the pack-ice-type rift areas. Thus, it is highly unlikely that these rifts would eventually lead to sea-floor spreading as assumed in some recent models of extensional orogenic collapse!

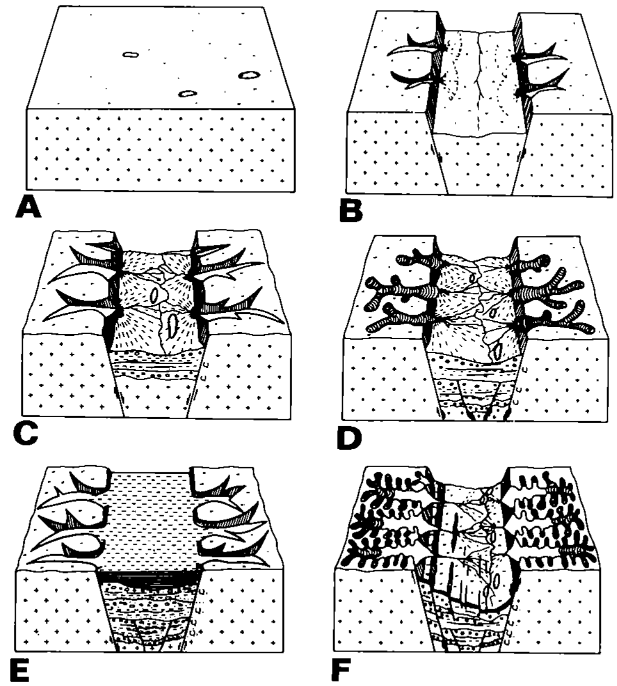


Figure 2.25 Schematic block diagrams illustrating geomorphological/sedimentary evolution of Tibetan intracontinental convergence-related rifts.
A. In late Pliocene, flattish erosional surface of Tibetan plateau was formed at an altitude of about 3 km (cf., Dewey et al., 1988; only 1 km according to Han et al., 1984!). Bedrock is shown by plus signs.
B. In early Pleistocene (and/or latest Pliocene), several N-S-trending extensional basins formed. Erosional gullies formed across their shoulders and alluvial deposits dominated by conglomerates began forming at fault scarps.
C. Accelerated deformation in southern part of plateau during middle Pleistocene is believed to have caused unconformity between Q_1 and Q_2. This was also a time of major glaciation. During interglacials, gullies further developed in newly created fault scarps.
D. Further tectonization in late Pleistocene created unconformity between Q_2 and Q_3 and generated further faulting and folding. The sudden rise of plateau by nearly 2 km led to creation of both valley and piedmont glaciers on the plateau.
E. In middle Holocene, advance of the last "interglacial" led to formation of numerous lakes, that in places, occupied glacially modified rift basins.
F. In later Holocene, evolution of mature Himalayan drainage and diminution of precipitation led to disappearence of many lakes. Normal faulting continued and glacial level in many places retreated to above 5,800 m. (Reproduced with permission from Han et al., 1984, Fig. 5.)

Rifts in intracontinental convergent belts get preserved only if the crust of the convergent belt is tectonically thinned *by an independent rifting event* immediately after the compression that created the rifts, to protect them from erosional destruction. This happened, for example, in the case of western Turkey, where north-south shortening created roughly north-south-trending extensional basins between the late Oligocene and the ?early Miocene, which basins were then preserved by rapid north-south extension that began sometime in the early Miocene (Şengör, 1987c, but with the revised timing of Seyitoglu and Scott, 1991). Another possible example is the preservation of the Permo-Triassic Ecca basins containing Karoo sedimentary rocks by the independent extension that led to the opening of the Indian and the Atlantic Oceans in the Mesozoic (Tankard et al., 1982, p. 380ff and Fig. 11–1). Rift belts related to intracontinental convergence do not seem to accomplish sufficient extension to thin the continental crust below its average thickness in order to take its top below sea level (they did not accomplish this in any of the examples just cited, and in very few other cases, the north European Zechstein basins being one remarkable exception: Ziegler, 1990). This is another reason for my skepticism concerning the reality of extensional orogenic collapse driven by the potential energy difference between thickened continental crust and its unthickened surroundings.

Pack-Ice-Type Rift Basins

As the name implies, the pack-ice-type rifts form from the jostling of blocks, into which fore- and hinterlands of collisional orogens become divided following the formation of impactogens, other collisional rifts and collision-related conjugate strike-slip faults. The best example of a pack-ice-type rift area is the group of rifts that formed and/or reactivated following the Adria/Europe collision along the Alps in the Eocene (Fig. 2.10; see also Illies and Greiner, 1976, Fig. 8; Dewey and Windley, 1988, Fig. 1) and that now forms the classical *Schollentektonik* of Germany and surrounding regions (cf., Stille, 1925; Lemcke, 1937; Lotze, 1937, 1971, p. 204–357; Martini, 1937; for a recent re-evaluation of the *Schollentektonik*, see Illies and Greiner, 1976; Meiburg, 1982).

The *Schollentektonik* in Germany and surrounding areas (Fig. 2.10), reaching from the Polish Trough (Pozaryski and Brochwicz-Lewinski, 1978) to eastern France (e.g., Burg et al., 1982) expresses a tectonic style characterized by the relative motion of rigid and semi-rigid blocks with respect to one another, along the margins of which compressive, distensive, and strike-slip deformation takes place (cf., Lotze, 1937, 1971, esp. p. 215). This tectonism is associated with dominantly alkaline magmatism (cf., Wimmenauer, 1974), and especially in the north, contains an important intracutaneous detachment horizon, namely the Zechstein salt (Lotze, 1953, 1957, 1971, esp. p. 216; Meiburg, 1982; Jaritz, 1987). In some places in northern Germany, salt-controlled detachment horizons also occur in the Triassic, and locally, the Zechstein and Triassic salt horizons may be mixed tectonically.

What created the multi-block structure shown in Fig. 2.10, and when, are still questions of dispute, but it seems that a general NW-SE-oriented broad right-lateral shear zone connecting the Carpathian area with the North Sea and reaching from the edge of the East European platform well into France and the Benelux countries is the most satisfactory explanation for the origin of these *Schollen* (e.g., Arthaud and Matte, 1977; see also W. H. Ziegler, 1975; for a different, but partly equivalent interpretation in terms of N-S compression, see Pratsch, 1982, esp. Fig. 12). The three major lineaments of the European basement, namely the Tempelburg-Kujawian uplift, the Fair Isle-Elbe Line fault zone, and the Danube-Pfahl-Frankian Line have functioned at least since the middle Jurassic as right-lateral shear zones[7] (perhaps since the Permian: see Arthaud and Matte, 1977; Pratsch, 1982; or even pre-Permian: see Ziegler, 1975, Fig. 6) and they were connected with each other through numerous cross elements oriented mainly NE-SW, compatible with tension-gash, conjugate-Riedel and rotated-tension-gash and conjugate-Riedel orientations (cf., Wilson, 1960; Tchalenko, 1970; see also Arthaud and Matte, 1977). Many of these cross elements have become extensional zones during the middle Jurassic to middle Cretaceous and accumulated thick strata. The main NW-SE-striking strike-slip systems also had, at this time, an extensional component. This extensional component along the main shear systems turned into compression in the Late

Cretaceous and remained so until the Eocene in many of them, extending well into the late Cenozoic in some. The NE-striking cross elements remained largely extensional through the Cenozoic, although a few of these also turned compressional (e.g., Martini, 1937; Meiburg, 1982). The cause of the Cenozoic (especially post-Eocene) events has been commonly ascribed to the Alpine collision in the south (Illies, 1974a,b; Illies and Greiner, 1976; Şengör, 1976b; Dewey and Windley, 1988). This collision reactivated many of the Mesozoic - earliest Cenozoic structures and created new ones within the larger blocks that had been already outlined by Mesozoic tectonics (e.g., the Filder Graben zone near Stuttgart or the Schweinfurt Graben north-northwest of Nürnberg: Fig. 2.10). The jostling of blocks created major grabens only in the Bohemian Massif (the Eger rift of late Oligocene age: Lotze, 1971, p. 190–193) and in the Rhenic Massif (the Lower Rhine Rift), plus the continuing rifting on the Upper Rhine impactogen. Smaller extensional structures, most likely only grabens, continued to extend along both sides of the Hessian depression (e.g., the Egge and the Leine grabens: cf., Meiburg, 1982) and farther north at the northwestern corner of the Subhercynian basin (Illies and Greiner, 1976).

The Rhenic and Bohemian massifs plus the South German block were internally disrupted elsewhere too, as indicated by alkaline basalts, phonolites, and other lavas. Small pull-apart basins also formed along the NW-SE-striking main shears in the Cenozoic, especially along the Fair Isle-Elbe line (Illies and Greiner, 1976).

The picture one obtains of the central European block mosaic in the Cenozoic is, then, one of basin formation in diverse orientations and through diverse mechanisms with variable magnitudes of extension across or indeed along them. Some basins pass along their trend into transpressional faults (e.g., pull-apart basins along the Fair Isle Elbe line), others pass into transtensional systems (e.g., grabens in the Hessian depression), and others have changed character during their evolution (e.g., Upper Rhine rift). In this mosaic, basins have formed at diverse times following the collision in the Alps from the Eocene to the Plio-Pleistocene! All of these observations can be most comfortably interpreted in terms of jostling in a loose-block mosaic driven in front of the main compressional front of the Alps, indeed resembling drifting pack ice.

Differences Between Collision-Related Rifts and Aulacogens

Both aulacogens and most varieties of collision-induced rifts are kinds of "high-angle rifts" with respect to the associated orogenic belts and/or ocean margins (cf., Şengör et al., 1978). For this reason, many of them have been mistaken for the other in many parts of the world (e.g., Wilson, 1954 mistook the Ottawa-Bonnechere aulacogen [cf., Burke and Dewey, 1973] for an impactogen; Burke, 1977, misidentified the Upper Rhine impactogen as an aulacogen), in most cases simply because the authors did not appreciate the differences between the two fundamentally different classes of rifts. In the following paragraphs, I briefly outline their differences to give the reader rules of thumb to distinguish the two groups.

In sharp contrast to aulacogens, collision-induced rifts do not have any stratigraphic relation to the ocean from which the orogen adjacent to these rifts originated. The pre-rift basement of collision rifts is made up of the rocks of the orogenic fore- or hinterland, and their rift fill is clearly post-collisional. Post-rifting doming will give rise to clastic wedges that, along with the correlative fanglomerates of the rift fill, would reveal the age of doming, as in the case of the Rhine rift and the Baikal rift.

The decisive test of whether a high-angle rift is an aulacogen or an impactogen would be the detailed study of the stratigraphy of its sedimentary fill: sedimentary fill correlative with the continental-margin assemblages would indicate that the rift is an aulacogen, whereas one correlative only with the molasse sequences of the associated orogen would suggest an impactogen.

However, unless the rift is subsequently deformed, its pre-rifting basement and the sedimentary fill of its rift trough are commonly not accessible to field examination. Even if the sedimentary record is revealed, either by drilling or by geophysical methods, it is commonly nonmarine, and therefore, presents difficulties for dating. Thus, identification in the stratigraphic record of the adjacent mountain belt opposite a high-angle rift of an exceptionally large delta and a deflection of the orogen toward the rift at this spot (e.g., the Vilyuy rift and the Verkhoyansk fold-and-thrust belt in eastern Siberia: Parfenov, 1984, Fig. 13; Gusev et al., 1985) may be used as another, possibly more practical, criterion for identifying the rift in question as an aulacogen.

Early, pre-rift doming and early syncollisional strike-slip movement parallel with the rift axis are additional, but risky criteria for the identification of aulacogens. Impactogens normally do not have associated pre-rift doming, and strike-slip deformation along the graben axis commonly begins late in their history, as shown by the history of the Upper Rhine rift.

Despite all these criteria, however, if ocean opening follows immediately a collisional orogeny and takes place parallel with the latter (e.g., as in the case of the Appalachians), then it may still be difficult to decide whether associated high-angle rifts belong to the collision phase and are thus impactogens, or to the ocean-opening phase, and are thus aulacogens. A further source of confusion would be if a rift originates as an impactogen related to one cratonic margin, and shortly thereafter is reactivated into an aulacogen by rifting along the opposite side. The geological history of the Dnyepr-Donetz aulacogen may be an example of such a "compound" case.

Intracontinental Wrench-Related Rift Basins

Brief History of Thought on Intracontinental Wrench-Related Rift Basins

The great Swiss geologist Arnold Escher von der Linth was possibly the first to have mapped a major strike slip fault in the Sntis Mountains south of the Wildkirchli in the Canton of Appenzell in the 1850's (Suess, 1883, p. 153–154). Suess (1883) was the first to note their widespread occurrence in mountain ranges. Kossmat (1926) and von Seidlitz (1931, esp. Chapters 40 and 41) were the first to imply that lateral motion of rigid or semi-rigid blocks may create extensional basins of significant size. Becker (1934) and Lotze (1937) explicitly showed how strike-slip motion may generate such basins by extension parallel with strike slip and by transtension, respectively. These two types of rift basins constitute the two main types of intracontinental wrench-related rift basins that we recognize today. That is why I prematurely end my historical survey of the evolution of thought on these basins here and continue the narrative in the next section.

Two Fundamental Types of Intracontinental Wrench-Related Rift Basins

Rift basins that form along strike-slip faults in relation to wrench faulting may be divided into two fundamental classes: 1) Basins that form along wrench zones with a zero component of extension across the zone (Fig. 2.26A), and 2) basins that form along wrench zones with a greater-than-zero component of extension across the zone (transtensional zones in Harland's, 1971, p. 30 terminology: Fig. 2.26B). Of these two, class 1 is really the only purely "wrench-related rift basins". Class 2 may be best described as "wrench-modified rift basins". As Fig. 2.26 shows, these two classes correspond with the basin styles described by Becker (1934) and Lotze (1937), respectively.

Wrench-Related Rift Basins

Wrench-related rift basins result from imperfections encountered in the strike of the wrench fault or fault zone. These imperfections cause local divergences between the strike of the fault or the fault zone and the slip vector. If the imperfection is orientated in such a way as to lead to a local extension along the fault or the fault zone, then it is called a "releasing bend" (Crowell, 1974b) and is contrasted with a "restraining bend", where the imperfection is orientated in such a way as to lead to local shortening (Crowell, 1974b).

All releasing bends give rise to pull-apart basins (Burchfiel and Stewart, 1966), across which the amount of extension equals the cumulative slip along the strike-slip fault since the origination of the releasing bend. Depending on the mode of formation of the releasing bend, the geometry and character of the pull-apart basins that form across them change.

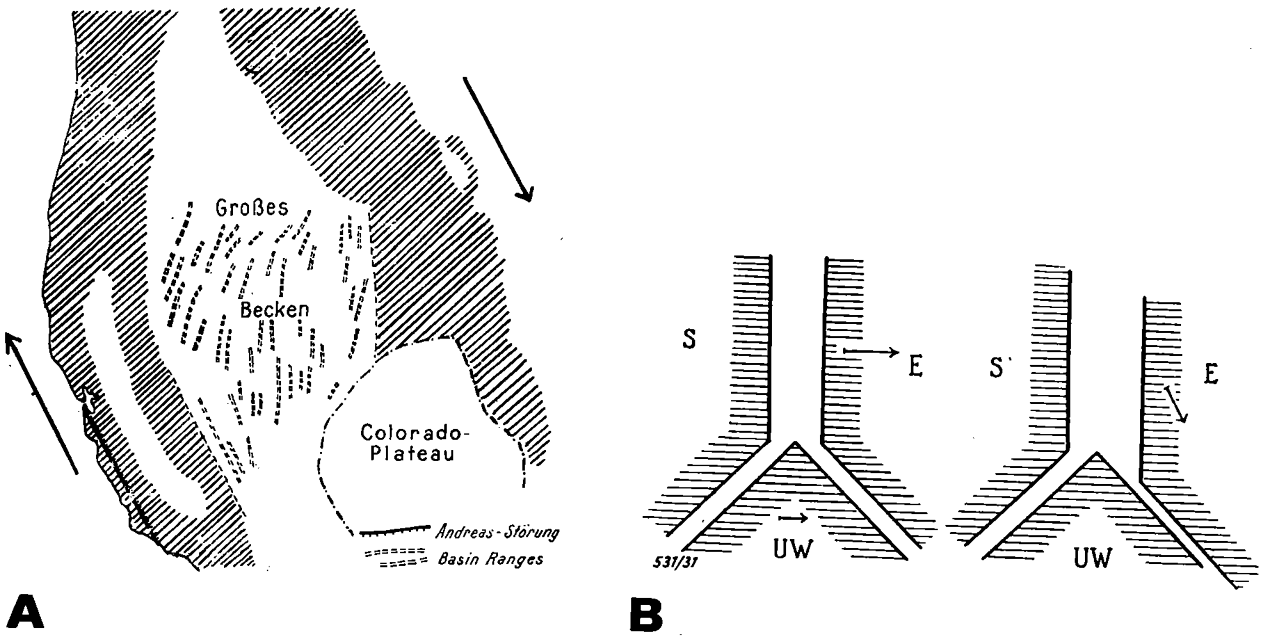


Figure 2.26 Two fundamental types of intracontinental wrench-related rift basins:
A. Basins that form along wrench zones with a zero component of net extension across wrench zone are illustrated here by Hans Becker's interpretation of San Andreas fault (*"Andreas-Störung"*)/Basin and Range region. As present Pacific/North America motion is essentially pure strike-slip (cf., Minster and Jordan, 1978, p. 5335), no net extension occurs *across* the shear zone depicted here, although enormous amounts of extension oblique and parallel to the shear zone have been observed in the Basin and Range province. (Reproduced from Becker, 1934, Fig. 3.)
B. Basins that form along wrench faults, across which there is a net component of extension (i.e., transtensional wrench; for location, see Fig. 2.10). The triple junction depicted is the Eichenberg graben knot in central Germany and illustrates the kinematics of a rift star. Key to abbreviations: S: Solling block; E: Eichsfeld block; UW: Unterwerra block.

Accordingly, in the following paragraphs, I discuss the tectonic and sedimentary relationships in wrench-related extensional basins in the framework of the various modes of formation of the releasing bends.

Restraining bends may also give rise to wrench-related rift basins, by tearing at high angle to the slip vector of one of the fault walls inhibiting motion. In such basins, the amount of extension equals the strike-slip separation since the origin of the basin, minus the shortening across the restraining bend immediately along the fault and reduces to zero away from the fault. Such basins form Carey's (1958) sphenocasms and their formation leads to rotation around vertical axes of major portions of fault walls along large strike-slip faults.

Extensional basins caused by releasing bends along strike-slip faults

Along strike-slip faults, releasing bends may be either of *primary* or *secondary* origin. Primary releasing bends are those that form along with a through-going strike-slip fault system. Holmquist (1932), Wilson (1960), Tchalenko (1970), Wilcox et al. (1973), Bartlett et al. (1981), and Naylor et al. (1986) discussed how a throughgoing strike-slip fault (or fault system) commonly develops (see also Sylvester, 1988). Figure 2.27 illustrates a general model that underlies the discussions cited above. A general tenet of this model is that before a throughgoing strike-slip fracture materializes, the sheared zone is deformed along a belt considerably broader than the ultimate strike-slip fault zone. This shear belt is deformed first by homogeneous straining in the area of the future shear belt (Tchalenko's, 1970, "pre-peak strength deformation"). If the sheared area has a vertically layered structure (as in any gently deformed or undeformed sedimentary basin), "en-echelon fold" trains may appear at this stage, with characteristic angles of 140° to 150° to the orientation of the shear belt (Fig. 2.27A; see Wilcox et al., 1973, Fig. 9, for photographs of experimental folds created in a layered clay cake). Just before the peak shear strength is reached, a series of shear fractures with the same sense of shear as the main shear zone appears. These fractures make average angles of 12±1° with the main shear zone and are called *Riedel shears*. At peak strength, the Riedels are rotated to a maximum of 16°. During this stage also, another set of shears, with a sense of shear opposite to that of the main zone, forms. Its members are inclined to the main shear zone at angles of 78±1°, and they usually form either simultaneously with or just before the Riedels and in materials with a lower water content than those in which this second set does not develop

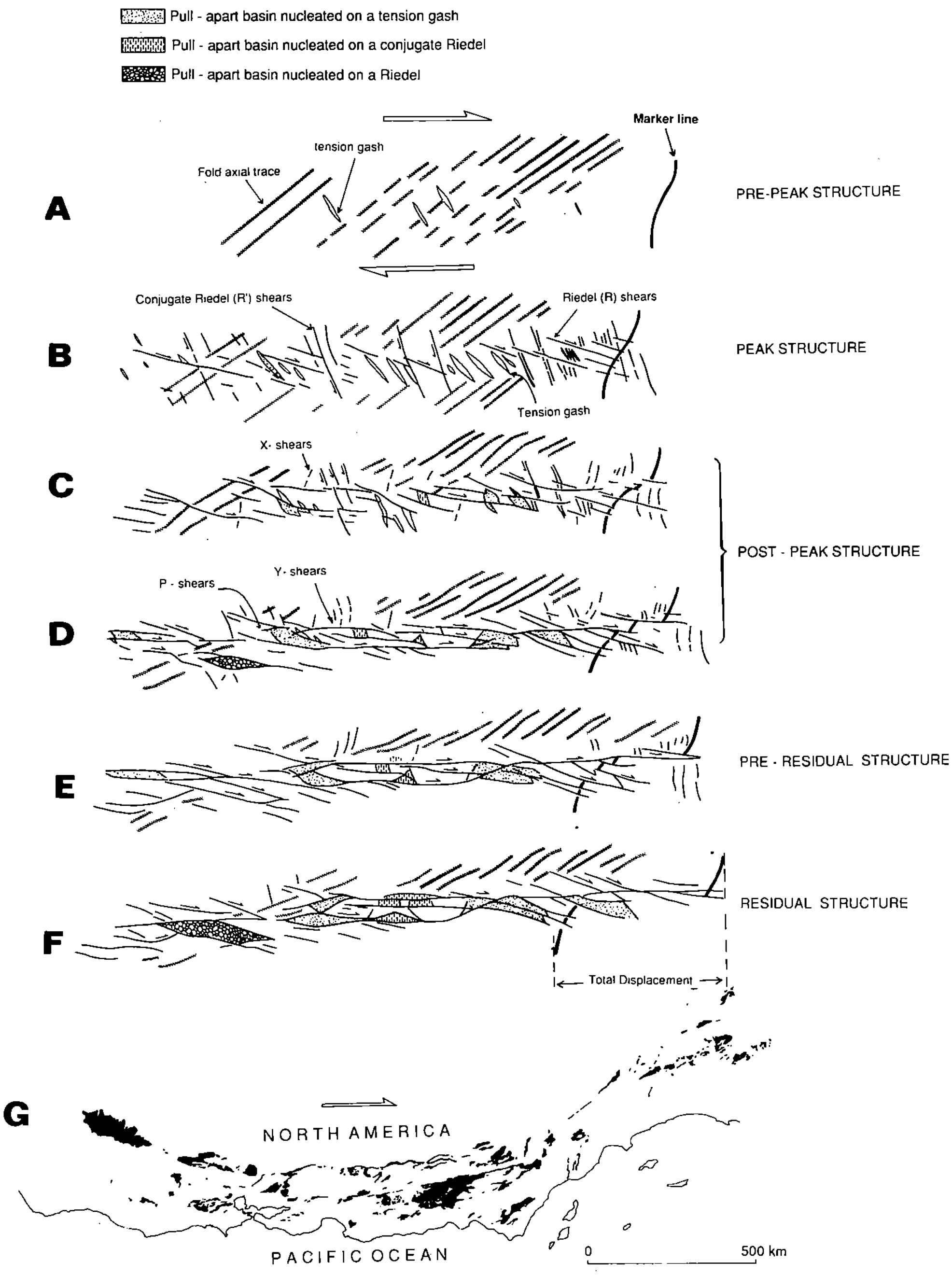


Figure 2.27 Sequence of structures, including pull-apart basins, that develop in a shear zone (**A** through **F**; see text for discussion). This sequence was developed on the basis of experiments and field work by Wilson (1960), Tchalenko (1970), Wilcox et al. (1973), Bartlett et al. (1981) and Naylor et al. (1986). Not all the structures had developed in any one experiment and the present figure illustrates "all possibilities". Notice how pull-apart basins form from openings of various shear-zone structures, such as tension gashes, Riedel shears, and conjugate Riedel shears, and grow to generate a string of basins. **A.** Pre-peak-structure stage; **B.** Peak-structure stage; **C.** and **D.** Post-peak-structure stages; **E.** Pre-residual-structure stage; and **F.** Residual structure stage. **G** shows a real geometry of wrench-related basin patterns from along the San Anrdeas system (however, showing both extensional and compressional basins! Reproduced with permission from the cover of Andersen and Rymer, 1983).

(Tchalenko, 1970). These are called the *conjugate Riedel shears,* which accomodate a maximum amount of shear between 0 and 50%. It is also at this stage that tension gashes may begin to open with strikes ideally orientated some 45° to the shear zone (Fig. 2.27B: Wilson, 1960) and with a direction of opening perpendicular to the strike of the tension gashes. When the peak strength is reached, both the Riedels and the conjugate Riedels are rotated to a maximum of about 16°, and also, some new shears form at angles of about 8°, along with the elongation of the Riedels with "flatter" tips (Fig. 2.27B). Tension gashes may continue to open and rotate into a sigmoidal form. The tips of such sigmoids may link up with Riedel shears and develop small pull-apart basins at this stage (Fig. 2.27B; see Wilson, 1960, esp. Fig. 3). The proportion of shear accommodated by these shears is about 75% (Tchalenko, 1970). At what Tchalenko (1970) termed "post-peak structure stage", an additional set of shears, called by him the *P-shears* form. They are usually formed at an angle of 170° (i.e., approximately symmetrically to the Riedels), and interconnect the latter, as seen in Fig. 2.27C. More than half of the shears are, at this time, inclined less than 4° to the shear zone and all the displacement takes place along the shears. The stage following this, what Tchalenko (1970) called the "pre-residual structure" stage, is characterized by formation of the first continuous horizontal principal displacement shears that isolate between them elongated lenses of essentially passive material (Fig. 2.27D).

It is clear from Fig. 2.27 that the principal displacement shears in a strike-slip zone come into being by interconnection of Riedels and P-shears, plus, in areas of incompatibility (i.e., where asperities form), by the additional generation of "flat" shears. A complex array of factors determines which particular Riedels and P-shears will link with one another to form the principal displacement shears (i.e., the throughgoing fault zones). These factors range from the anisotropies in the failing medium (both material-and structure-governed) through the displacement rate (Hempton, 1983) to the state of stress before wrenching begins (e.g., Naylor et al., 1986).

Figure 2.27D shows how both releasing and restraining bends may come into existence together with the main displacement shears, being nucleated on former Riedel shears or tension gashes and more rarely on the conjugate Riedels. Depending on the structure on which they nucleate, the resulting pull-apart basins will have various geometries and initial overlap distances of the master faults (*sensu* Rodgers, 1980, Fig. 3; Segall and Pollard, 1980, call the negative overlap *separation,* and they call *step,* what Rodgers, 1980, calls separation; it is Rodgers' terminology I follow here). In a numerical study, Rodgers (1980) has shown that the overlap and separation of the two master faults bounding a pull-apart basin are important parameters in determining vertical motions of the ground and the orientation of faults that will form around the ends of the principal faults. They therefore determine the geometry of the pull-apart basin and the topography of its surroundings.

One important aspect of Rodgers' models is division of the locus of extension into two nodes as the pull-apart basin lengthens parallel with the slip along the master faults. This is likely to happen in all pull-apart basins, as soon as the master-fault overlap equals the fault separation. Thus, in most well developed pull-apart basins, we expect to find two separate nodes of subsidence at the ends of the basin, separated by a "central horst", if the models of Rodgers (1980) apply.

Few pull-apart basins have been investigated in sufficient detail to see whether Rodgers' (1980) prediction is fulfilled. Rodgers (1980) was unable to cite any examples to corroborate his model with two loci of subsidence at the ends of a pull-apart basin. Both active and fossil examples of this situation, however, are known. The Dead Sea pull-apart basin now has two sub-basins, separated by the Lisan Peninsula swell (see Neev and Emery, 1967, Fig. 8, for a bathymetric map of the Dead Sea) and this geometry also existed during the Pleistocene, when the Dead Sea was occupied by the larger pluvial Lake Lisan (Neev and Emery, 1967). The similarly active Vienna pull-apart basin, a "thin-skinned" structure, also has two loci of maximum subsidence, located at both ends of the depression bounded by two major strike-slip faults (Royden, 1985, esp. Fig. 14). Finally, the Jurassic-Cretaceous fossil Soria basin in northern Spain has two loci of maximum sediment thickness at both its ends, although they are not exactly coeval.

The above considerations give us the first clues as to what sedimentation in a primary pull-apart basin ought to be like. For negative or null overlap, no high ground is created outside the pull-apart basin, unless it be a result of small fold development at the ends of the main fault strands (cf., Segall and Pollard, 1980, esp. Figs. 9d and 11d). The low ground that forms as a result of extensional faulting in the separation area has an "island" of no extension in the case of negative overlap (Segall and Pollard, 1980, esp. Figs. 9b and 11b). But as the main faults propagate to reach a condition of zero overlap, that "island" is eliminated and a zone of subsidence is created with a long axis connecting the tips of the propagating main shears (Rodgers, 1980, Fig. 4d). At this stage, still no high ground is necessarily created outside the pull-apart basin.

Figure 2.28 displays the predicted sedimentation patterns and the distribution of facies in a primary pull-apart basin. In the negative-overlap stage (Fig. 2.28A), normal faulting leads to clastic sedimentaton in the basin that is fed both from the basin edges and from the internal "island". Absence of highlands around the basin precludes prediction of preferred provenance sites for the clastics, which are functions of local geomorphic and climatic factors.

By the time the zero overlap stage is reached (Fig. 2.28B), the internal clastic source (the "island") is eliminated and the basin is fed only peripherally. At this time, the "ends" of the basin (i.e., the sides not delimited by the main shears) are the most likely avenues of sediment intake, because they lie lower than the "sides" of the basin (i.e., the edges of the basin bounded by the main shears) (Rodgers, 1980, Fig. 4d).

As overlap of the main shears reaches a value of 1/2 of the separation (Fig. 2.28C), two ridges trending almost perpendicularly away from the tips of the main shears begin rising, as the basin is elongated between the propagating main shears, still with a central depocenter. These ridges likely become sediment sources, and thus increase the clastic flux into the pull-apart basin and probably also the grain size of the incoming clastic sediments.

Further elongation of the pull-apart basin by increasing offset along the master shears results eventually in the overlap reaching a value equal to the separation (Fig. 2.28D). In this case, extension is localized at the "ends" of the pull-apart basin, creating two distinct nodes of subsidence that commonly divide the pull-apart basin into two "sub-basins". These sub-basins continue to be fed by the erosion of the highlands that trend away at high angles to the main shear zone from the tips of the master shears and may temporarily protect the center of the pull-apart basin from the axially fed sediment influx, in a way entirely reversing the situation at the "zero overlap" situation. A simple corollary to this is that, in an elongate pull-apart basin (Fig. 2.28E), a uniform total sediment thickness in the basin may not necessarily imply a layer-cake stratigraphy! As Rodgers (1980, Fig. 7) showed, further elongation of the primary pull-apart basin simply accentuates the situation attained by making separation equal to overlap.

Secondary releasing bends along a strike-slip fault form following a finite interval of time after a throughgoing fault strand is established. How this comes about has been investigated by Merzer and Freund (1975), J.F. Dewey (unpublished, 1979), Mann et al. (1983), and Şengör et al. (1985). The following is a brief synopsis of the main mechanisms for the formation of secondary releasing bends:

1) Strike-slip fault "buckles" during motion as a consequence of the compressional buckling of its bounding crustal boards (Merzer and Freund, 1975). This phenomenon commonly affects a family of parallel strike-slip faults, separated from one another by crustal boards of similar width (Merzer and Freund, 1975).

2) Strike-slip fault is intersected by a zone of high convergent and/or divergent strain. The segment affected by the intersection may be rotated away from its initial strike and thus may change its orientation with respect to the slip vector, causing either a releasing or a restraining bend (Mann et al., 1983, Fig. 3, mechanism "C"). Constriction of an escaping continental wedge from a zone of convergence (Mann et al., 1983, Fig. 3, mechanism "D"; e.g., the Anatolian Block or Scholle escaping from the Turkish-Iranian High Plateau convergent zone: Şengör and Kidd, 1979; Şengör et al., 1985) is a special case of such a situation. Another special case is

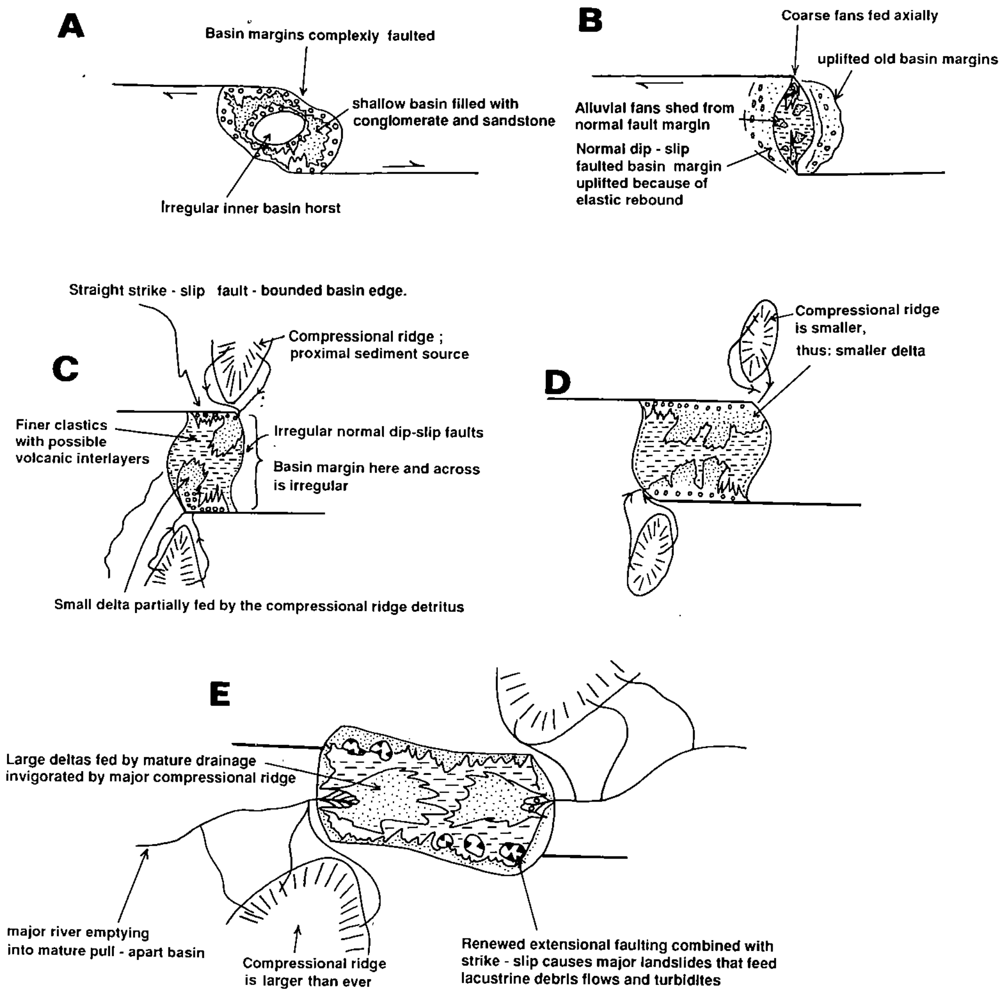


Figure 2.28 Evolution of sedimentary facies in a developing pull-apart basin, located in the temperate zone and in dominantly non-calcareous terrain.

A. Separation is 1/2 of negative overlap. Notice here that an "insular," unextended region functions as an inner-basin horst, providing intra-basinal sediment.

B. Zero overlap. This is equivalent to an "opening" conjugate Riedel shear.

C. Separation is twice the overlap. Note enlargement of highlands outside pull-apart basin and corresponding increase of influx of clastic sediment into the basin, commonly building deltas and subaqueous fans emanating from the two basin corners facing the highlands.

D. Separation equals overlap. The highlands are diminished, but they still supply material to commonly subaqueous fans in the basin, whose stratigraphy would exhibit an upward fining sequence with respect to the last frame.

E. Separation is twice the overlap. The relief between basin floor and highlands has increased dramatically compared with the last frame and normal faulting has begun invading territory outside bounding faults. Increased relief may increase clastic influx, and aided by juvenile normal faulting disrupting steep bounding master fault walls, may begin triggering massive landslides. *All this is most likely to be interpreted in sedimentary/structural record of pull-apart rift as a "phase of renewed rifting," wheraes in reality no increase in rate of strike-slip faulting is introduced.* This would thus give rise to an interpretation in terms of episodic tectonism, while in reality tectonism is perfectly uniform, but not its record!

the rotation of a strike-slip fault away from parallelism with the slip vector along its original course in a general area of quasi-homogeneous shortening as, for example, in Central Asia (Şengör, 1987d, Fig. 22) or in the Turkish-Iranian High Plateau (Şengör et al., 1985, Fig. 10). In such a situation, the original strike-slip fault may be broken into a series of pull-apart wrench-fault segments (mechanisms "A" and "B" in Fig. 3 of Mann et al., 1983).

3) Asperities along a strike-slip fault may cause local rotation of the strain field and create splay faults which begin to accommodate motion. These splay faults are at angles greater than zero degrees to the slip vector and may thus nucleate restraining and/or releasing bends (J.F. Dewey, personal communication, 1979).

The pull-apart basins that will form along secondary releasing bends are in principle, no different from those that form along primary releasing bends, except for the angle their long axes make with the slip vector. This angle is, in principle, more variable in the case of pull-apart basins forming along secondary releasing bends because these bends do not necessarily nucleate on structures that have fixed orientations with respect to the orientation of the slip vector, such as tension gashes or Riedel shears.

Extensional basins formed at restraining bends along strike-slip faults

Sphenochasms are the only major type of extensional basins that form at restraining bends along strike-slip faults. Sedimentation in such basins ranges from entirely continental to oceanic, depending on the amount of extension. Commonly, total thicknesses of sedimentary rocks decrease away from the major bounding fault, as seen, for example, in the case of the northwest Chinese graben complexes (Fig. 2.18: Ma et al., 1982; Burchfiel and Royden, 1991).

If no other factors interfere, then major relief in sphenocasms forms near the bounding strike-slip fault, both because of this faulting, and because it is near the bounding strike-slip fault that the amount of extension (and the presumably resultant shoulder uplift) is greatest (e.g., as in the Rio Grande sphenochasm: Baltz, 1978, esp. p. 211; Chapin, 1979, p. 3; see also Ingersoll et al., 1990). This results in coarse clastics being deposited where the basin is widest and the grain size diminishes progressively away from the main bounding strike-slip fault (see for example the "Late Tertiary Palaeogeography of China" on p. 126 in Wang, 1985, for a map of the Northwest China Graben System showing a dimunition of grain size towards the north).

Wrench-Modified Rift Basins

Wrench-modified rift basins are those extensional basins in which extension occurs in an oblique orientation, imposing on the basin also a strike-slip component. Although many pull-apart basins and sphenochasms may be included in this definiton, I intend to exclude them by imposing a further restriction on the definition of wrench-modified extensional basins, viz. that the main tectonic grain in which they form must be parallel with their long axes. This excludes the presence of major strike-slip systems with strikes oblique to the basin, along which the basin may have opened as a pull-apart. In my sense, wrench-modified extensional basins are equivalent to Harland's (1971, p. 30, esp. Fig. 2) *transtensional* basins and also correspond with *leaky* strike-slip faults.

As I point out above, most rifts have transtensional components. However, when transtension becomes the characteristic regime of an extensional basin, it develops certain tectonic features peculiar to this environment that influence sedimentary processes and resultant sedimentary rock bodies. Figure 2.29A illustrates a simplified block diagram of a transtensional basin, in which normal-fault trains oblique to the main basin margins resemble tension gashes that form in strike-slip environments. Such oblique trends dominate many transtensional basins (e.g., Gulf of California: Kelts, 1981, 1982; Sea of Marmara: Şengör et al., 1985; Barka and Kadinsky-Cade, 1988) and the following discussion on the sedimentation and tectonics in such basins is mainly based on examples drawn from the now well exposed Central High Atlas basin in Morocco (Beauchamp, 1988; Laville, 1988).

Beauchamp (1988) and Beauchamp and Petit (1983) have shown, following earlier authors (e.g., Dewey et al., 1973; Van Houten, 1977; Manspeizer et

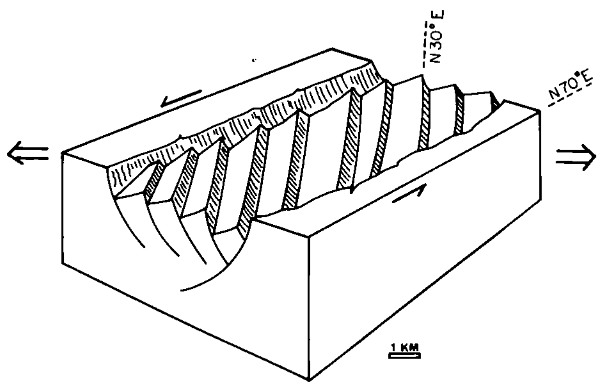
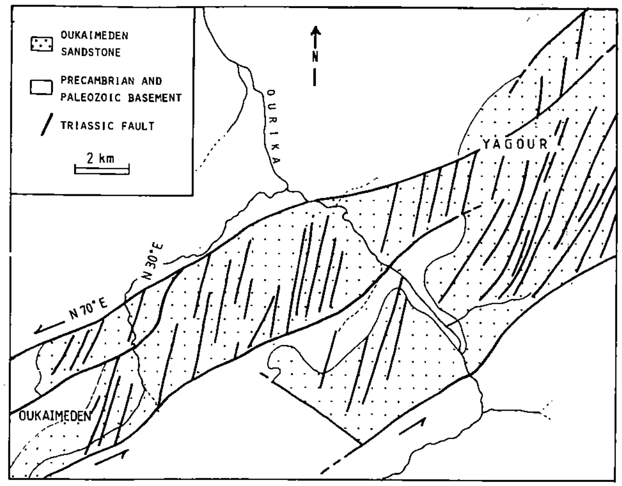


Figure 2.29 Geometry of faulting in transtensional basins:
A. A schematic block diagram illustrating the pattern of faulting in fossil intracontinental transtensional rift basin on the basis of data gathered from the Esçour basin in the High Atlas of Morocco. (Reproduced with permission from Beauchamp, 1988, Fig. 20-10.)



B. Pattern of Triassic faulting in a fossil transtensional intracontinental rift located in the Ourika region of the High Atlas of Morocco. (Reproduced with permission from Beauchamp, 1988, Fig. 20-11.)

al., 1978; Şengör et al., 1984), that the Triassic-Jurassic geological history of northwest Africa was dominated by transtension that detached the largely Hercynian deformed Moroccan and Oran mesetas from the Precambrian Saharan platform (Fig. 2.30). The present structure of the Central High Atlas mountains reflects the geometry following early Cretaceous through Cenozoic compressional deformation that moderately to gently overprinted Permo-Jurassic extensional structures. As Beauchamp (1988) and Laville (1988) have pointed out, following earlier authors, the structure of the Central High Atlas mountains is dominated by basins and ridges oriented obliquely to the main ENE grain of the mountain chain. Beauchamp (1988) has shown that the main oblique features correspond with NNE-trending ridges formed as normal-fault-bounded horst blocks or half horsts, as displayed in Fig. 2.30. Laville (1988) has shown that extensional structures form a more complicated network in the transtensional basin than the simple diagram in Fig. 2.29 would lead one to suspect. He distinguished three main kinds of oblique "ridges" that formed during transtension:

1) Extensional ridges. These ridges trend NNE to NE and consist of two distinct varieties:

1a) Extensional ridges without intrusions. These are commonly monoclinally tilted blocks or two-faced horsts bounded by normal faults. Some show a sigmoidal outline, ending at NNE-striking strike-slip faults (many now turned into steep thrusts) and others splay away from such faults.

1b) Extensional ridges with intrusions. These are anticlinal structures cored by early Mesozoic intrusions of gabbros, leucogabbros, and syenites in some places forming simple dikes and in others, ring complexes . In map view, they are sigmoidal to lozenge-shaped and appear to have filled in tension gashes opened as a result of transtension.

Einsele (1982, 1986) has developed a model to elucidate the evolution of such magmatic centers in transtensional systems and their relationship with basin sediments, which is illustrated by Fig. 2.31. Einsele's model shows that upwelling basaltic magma in pull-apart basins or in small spreading ocean basins in transtensional environments (e.g., the Guaymas basin in the Gulf of California: Curray et al., 1982) not only creates dike complexes or large intrusive bodies, but also sills in soft sediment accumulating in such basins. The thickness of the sill-sediment package is controlled chiefly by the rate of sedimentation. Extremely high sedimentation rate surpassing the spreading rate causes filling up of the basin and hampers the rise of magma to the surface (as in much of the High Atlas Mountains). Very low sedimentation rate (e.g., those prevailing at the major mid-oceanic

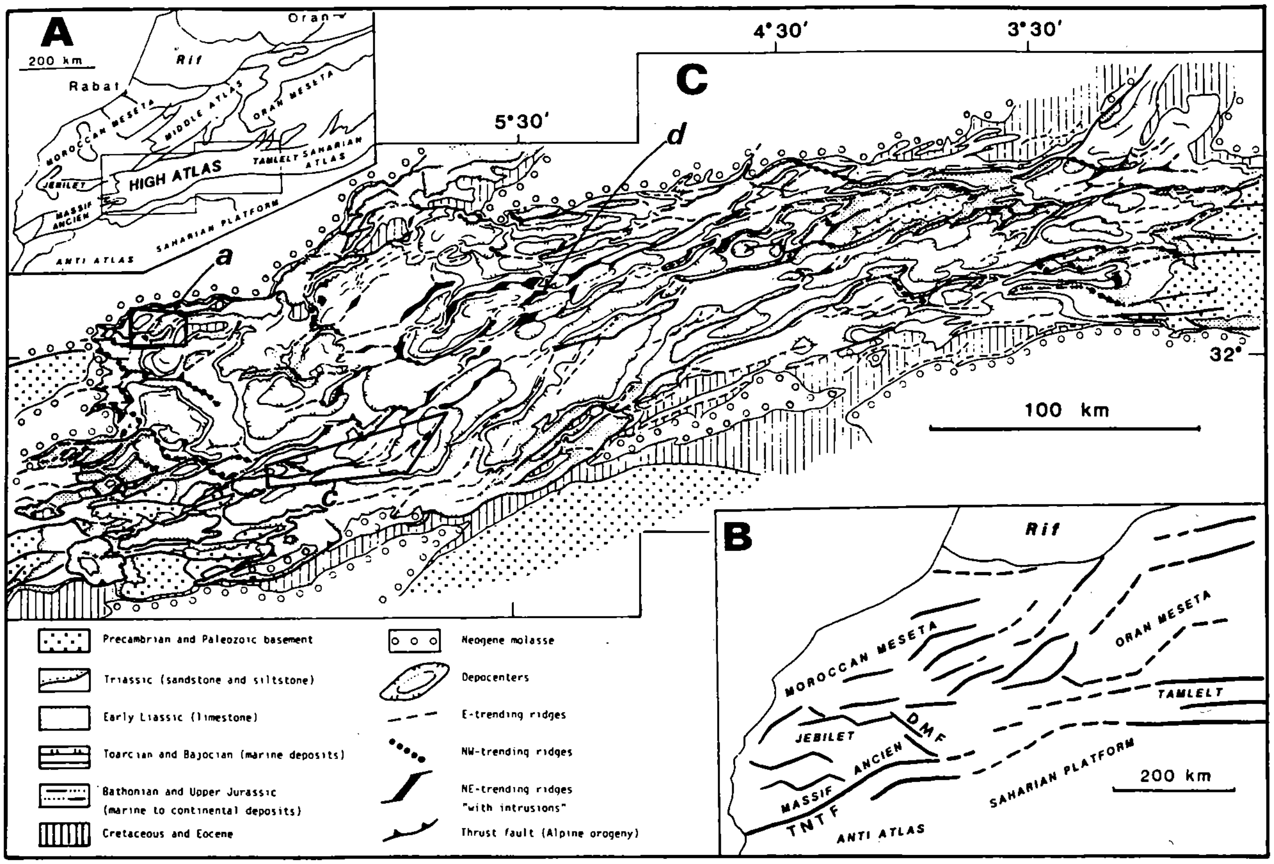


Figure 2.30 The present structural picture of the Central High Atlas Mountains, Morocco (reproduced with permission from Laville, 1988, Fig. 21-1,): **A.** Tectonic units of Morocco; **B.** Late Hercynian fault pattern in Morocco that has influenced nucleation of Alpide faults. (DMF = Demnat fault; TNTF = Tizi n'Test fault.) **C.** Simplified geological map of the Central High Atlas Mountains. Note on this map the spatial relationships between NE-trending "extensional" ridges with NW-trending "compressional" ridges, betraying a left-lateral strike-slip regime, even after intense N-S-directed Alpide shortening in Late Cretaceous and early Cenozoic.

spreading centers) hardly influence magma extrusion onto sea-floor. Intermediate rates that do not necessarily fill up the troughs may still generate sill-sediment thicknesses of up to a few-hundred meters (Einsele, 1986). In such basins, "sea-floor spreading" may not create conventional ocean floors, but instead generate what is commonly known under the various designations such as "slate-diabase complexes" (e.g., in the Greater Caucasus: Khain, 1975; Beridze, 1984) or "diabase-phyllitoid complexes" (e.g., in the Strandja Mountains in SE Bulgaria: Chatalov, 1991 and the literature cited therein) or indeed "schistes lustres" (Isler and Pantic, 1980) in the literature. The controversy created by the interpretation of Dewey et al. (1973) of the Atlas Mountains in terms of a transtensional "ocean" although no "true ophiolites" have yet been found in them, likely has resulted from just such a situation. Many of the slate-diabase, or diabase-phyllitoid, or schistes lustres basins in the central and western Tethysides have been interpreted in terms of transtensional troughs (e.g., the Pennine realm in the Alps: Kelts, 1981; the Slate-Diabase Zone in the Greater Caucasus: Şengör, 1990b, 1991c); these "sill-sediment" packages appear to be a hallmark of transtensional basins in general.

2) Compressional ridges. These ridges trend NW and appear either as anticlinal folds or thrust-bounded uplifts (Fig. 2.30). They are commonly smaller and less common than extensional ridges. Like the latter,

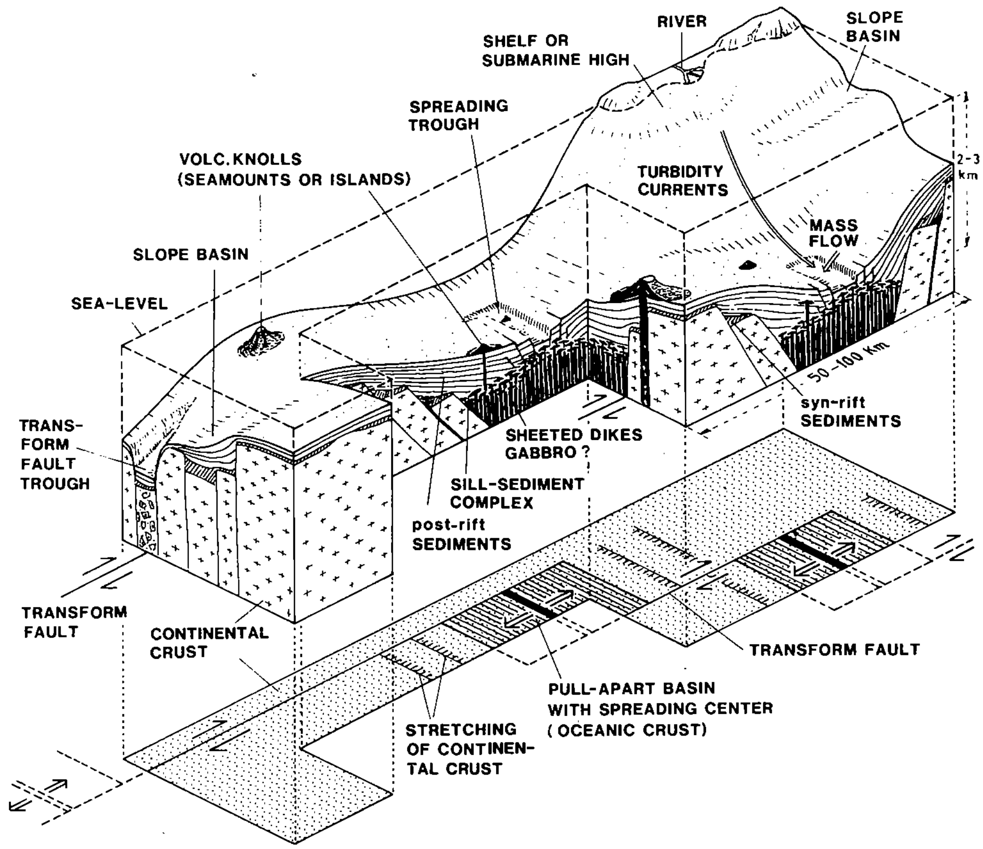


Figure 2.31 Einsele's (1986) model of pull-apart magmatism in transtensional systems. The top shows schematic block diagram, and the bottom shows the plan-view of Gulf of California-type basins (shown as simple pull-aparts for simplification; reproduced with permission from Einsele, 1986, Fig. 3).

they end on ENE-striking former strike-slip faults or splay away from them.

Numerous ENE-striking former strike-slip faults were active during Triassic-Jurassic transpression along the High Atlas Mountains and these acted as relay features connecting oblique elements with one another. These large ENE structures also had a normal dip-slip component (Beauchamp, 1988), reflecting the transtensional component of motion in the opening of the High Atlas basins, as do the considerably higher number and larger size of the extensional oblique ridges and basins than the compressional ones.

All of these structures were growing during Triassic-Jurassic extension, as shown by 1) thickness changes across them, 2) resedimentation by sliding material off their crests and flanks during the early and medial Jurassic (Lee and Burgess, 1978), 3) lateral facies changes between basins and ridges, indicating active uplift during deposition, and 4) local unconformities. Within the main trough of the Central High Atlas basin, all of these features were sediment sources (Beauchamp, 1988), although internal dynamics of the depositing medium is likely to have controlled the dominant NE to SW dispersal pattern observed by Beauchamp (1988).

The main sedimentary rocks deposited during earlier phases of rifting in the Triassic in the High Atlas near Marrakesh show a progressively greater influence of marine environments, ranging from entirely terrestrial alluvial-fan/playa-lake deposits through shoreline, delta, and lagoonal deposits. During much of the Early and Medial Jurassic, sedimentation produced a thick sequence of limestone and marl (in places up to 7 km thick). In the later Early Jurassic and earlier Medial Jurassic (Aalenian to Bajocian), deposition changed gradually from deeper-water marl to shallower-water bioclastic and oolitic limestone, punctuated by coral-algal reef horizons (Laville, 1988); in the Bathonian, the basin entirely filled up, leading to fluvio-deltaic sediment deposition (DuDresnay, 1979; Jenny et al., 1981). Harmand and Laville (1983) have shown that greenschist metamorphism of the sedimentary rocks (cf., Studer, 1980) shows concentric zonation away from a potassic center through an argillitic mantle to a prophyllitic periphery, and interpreted these as resulting from hydrothermal processes generated by a fossil geothermal field, similar to those seen in active and ancient fracture zones in the oceans (e.g., DeLong et al., 1979) or in the transtensional Gulf of California basin (e.g., Kelts, 1981; Einsele, 1986).

Discussion

The main purpose of this chapter is to show the tremendous variability in processes of rifting and to provide hints to help the reader to identify ancient rifts preserved in various states of completeness in the geological record. So far, emphasis in the literature has been placed mainly on individual rifts, rather than on systems of rifts. Even when large areas of rifting are covered by different authors, the general tendency has been to talk either about overall processes of stretching of the lithosphere, or about the structure of individual rifts. The main controversies concerning rifting lately have been concentrated either on whether rifting is an asymmetric process, creating a dominantly monoclinic structural fabric or a symmetric one with an orthorhombic fabric, or whether uniform or heterogeneous stretching best accounts for rifts. These polarizations of opinions have commonly come about between individuals or groups working on different rift systems or on different parts of the same rift system. Only later in the history of these controversies, the proponents of the two views have "invaded" a common ground and proposed radically different interpretations of the same tectonic objects.

To avoid falling into the trap of being blinded by peculiarities of any local structure, the first part of this chapter considers rifts as systems, as parts of larger structural entities called *taphrogens*. Such consideration has been common in the case of convergent structures, when they have been grouped into orogenic belts (or "folded belts"). That this has not occurred in studies involving regions of extension was mainly because taphrogens are commonly buried under their own fills, in contrast to orogens that stand high and naked of sedimentary mantle, providing a much more attractive target for the hammer of the geologist. Indeed, when J.W. Evans took "regions of tension" as the topic of his presidential address to the Geological Society of London in 1925, he complained that "The results of compression have long been studied, but comparatively little attention has been given to the occurrence of tension in areas where it has left evidence of its existence in the form of joints, normal or slip faults (occasionally replaced by monoclinal folds), or of dykes and other characteristic igneous phenomena" (Evans, 1925, p. lxxx).

Once we recognized that rifts occurred in organized families, we had the need to distinguish those "thick-skinned" extensional structures from "thin-skinned" ones. We employed the historical evolution of the two common terms, namely *graben* and *rift*, to aid us in this and we found it convenient, and consistent with historical usage, to restrict the word graben to thin-skinned extensional structures and to use rift for thick-skinned ones. Taphrogens, like orogens, include both thin-skinned and thick-skinned structures, so they consist of both grabens and rifts.

Our next task in our pursuit of the tectonics of extensional regions was a classification of rifts. This we did from three, mutually not completely overlapping viewpoints. We first classified rifts in terms of the plan view of the taphrogens or parts of taphrogens in which they occurred, in other words from a purely *geometrical* viewpoint. Then, in order to elucidate their *kinematics*, we made use of the framework provided by plate tectonics and distinguished intraplate, plate boundary, and triple-junction-related rifts. Finally, from a *dynamic* viewpoint, we were able to distinguish only two classes, namely rifts formed through the active participation of the mantle through convection-driven hot jets, called mantle plumes underlying hot spots on the earth's surface, and those that form by stress systems generated by the two-dimensional evolution of the world plate mosaic, in which the mantle plays only a passive role.

The purpose of our classification is to make an overview of the tremendous variability in the world's rift population easier, but this classification should not lead us to think that nature rigidly operates in its confines. As Stephen Jay Gould recently wrote "Classifications are *theories* about the basis of natural order, not dull catalogues compiled only to avoid chaos" (Gould, 1989, p. 98, italics mine). My classification also is a theory of rift origin, or, rather, a group of theories nested one within the other. It is proposed to allow us to recognize not only ancient individual rifts but also whole taphrogens even when only small parts of them or even parts of their constituent rifts and grabens are preserved. Figure 2.5 may be thought of as a flow chart within the framework of this identification process. In the text, I frequently stress the formation path of a rift or a taphrogen as a whole, when discussing its origin, for not *every* combination of items listed on this table seems allowed and this may aid us in filling in the missing parts from the records of ancient rifts.

I must stress the importance of recognizing the *continuous spectrum* that exists between the end members displayed on Fig. 2.5. Nature does not know about my classification scheme: the classes of this scheme *grade* into one another in nature and only rarely form well defined pigeon holes. Although pigeon holes are necessary for creating pedagogic aids and a structured history, rational thought demands continuity of process. It is within this consciousness of the continuity of process that the classes of this classification are to be understood only as aids to comprehension and retention.

Our scheme of classification guides us through a tour of the three most common types of ancient rifts encountered in the archive of historical geology, namely, aulacogens, collision rifts, and intracontinental strike-slip-related rifts. In each case, we find it useful to begin our discussion with a brief historical overview to enable us to see what sort of biases may exist in the handling of the specific problems we are to tackle. We then deal with the various representatives of the three types of ancient rifts, and in this we try, as much as possible, to keep a truly global perspective. The emphasis has been on what there is rather than what there ought to be, following implicit advice given to students of rifts by Warren Hamilton (1985a, p. 362).

Tectonic environments in which aulacogens, collision rifts, and intracontinental wrench-related rift basins form have been considered in greater detail than in most previous discussions of rift origin, and even among these three groups, *transitions* have been noted. In rift studies, especially in considering the historical geology of rifts, processes ought to be thoroughly considered and the record seen in rifts should be interpreted, as much as possible, in the light of processes of which we have first-hand knowledge (i.e., presently active processes). *The history of rifts does not consist only of the record preserved,* which forms only a fraction, and commonly a *small* fraction of the total history. To get the entire history, we must *interpolate* between record fragments and that can only be done soundly in the light of *actualistic* thinking.

The processes to be considered make up a spectrum. It ranges from purely structural ones, creating the rift valleys through igneous and sedimentary processes involved in filling the rifts, to weathering and erosion that destroy the rift. Such seemingly remote processes as the orbital controls on climate have recently opened up new and hitherto undreamed-of high precision in dating and correlating sedimentary contents of rifts. Similarly, hydrothermal-circulation studies along master faults have revealed thermal anomalies in the evolution of ritfs not previously considered. All such truly multi-disciplinary studies on ancient rifts tell us that rift studies remain an exciting field that carries much promise for new and ground-breaking advances. This acquires added significance when we consider that Coppens' "east-side story" holds the formation of the east African rifts system directly responsible for generation of the hominids, and thus, of humans (Coppens, 1991)

Acknowledgments

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Endnotes

[1] I must emphasize that I am very skeptical about "extensional orogenic collapse" under the weight of uplands alone. Everywhere it has occurred, an additional process, such as tectonic escape, seems to have aided it. Moreover in many places where it is proposed to occur, the number of structures responsible for extension and the actual amount of stretching are far smaller than proposed (e.g. the Alps), and the direction of stretching is at variance with the solely gravity-driven extensional orogenic-collapse model (e.g. the Betic Cordillera).

[2] Jarrard's (1986a) seven or even five different types of arc behavior are far too detailed to be applicable to a general survey of the historical geology of arcs and are therefore not used here. For a discussion, see Şengör (1990a, p. 66).

[3] A synclise is essentially a commonly equant basin in the Soviet terminology: see Dennis (1967, p.147–148) for details; Shatsky even used it for what he later called aulacogens!

[4] Shatsky originally said "syneclise", but because most of his syneclises since have turned out to be either oceans (e.g. the North Caspian "syneclise": see Burke [1977a]) or flexural foreland basins, there is no danger in exchanging "major basin" for the "syneclise" in the definition.

[5] For rift propagation in environments of continental separation, see esp. Courtillot (1982),Courtillot and Vink (1983), and McKenzie (1986); for a present-day example of a propagating rift, see the Sadko trough and its southern continuation in the Chersky Mountains where the Balagan Tas volcano and active normal faulting indicate that the Gakkel spreading center is propagating into Asia: Vogt and Avery (1974), p. 104ff and Fig. 14; Fujita and Cook, 1990, esp. Fig. 2

[6] Known since the beginning of the nineteenth century, when folklore referred to them as "the footprints of Noah's Raven" (see Colbert, 1968, p. 37ff).

[7] These major lineaments are not everywhere expressed as single fault lines. Especially in the north German/Polish basement, they are rather broad zones of complicated faulting and some folding, resulting from the distribution of shear in a broader zone than in the south, and thus do not show up on seismic-reflection lines. This should not, therefore, be interpreted as their being absent in the north!