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Notes

Global catalogue, classification and tectonic origins of restraining- and releasing bends on active and ancient strike-slip fault systems

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Abstract: Restraining- and releasing bends with similar morphology and structure have been described by many previous studies of strike-slip faults in a variety of active and ancient tectonic settings. Despite the documentation of at least 49 restraining and 144 releasing bends along active and ancient strike-slip faults in the continents and oceans, there is no consensus on how these structural features are named and classified, or how their wide range of structures and morphologies are controlled by the distinctive strike-slip tectonic settings in which they form. In this overview, I have compiled published information on the strike-slip tectonic setting, size, basin and bend type, age, and models for active and ancient releasing and restraining bends. Examples of bends on strike-slip faults are compiled and illustrated from five distinctive active strike-slip settings:

1. oceanic transforms separating oceanic crust and offsetting mid-oceanic spreading ridges;
2. long and linear plate-boundary strike-slip fault systems separating two continental plates whose plate-boundary kinematics can be quantified for long distances along strike by a single pole of rotation (e.g. the San Andreas fault system of western North America);
3. relatively shorter, more arcuate indent-linked strike-slip fault systems bounding escaping continental fragments in zones of continent–continent or arc–continent collision (e.g. the Anatolian plate);
4. straight to arcuate trench-linked strike-slip fault systems bounding elongate fore-arc slivers generated in active and ancient fore-arc settings by oblique subduction (e.g. Sumatra); and
5. cratonic strike-slip fault systems removed from active plate boundaries, formed on older crustal faults, but acting as ‘concentrators’ of intraplate stresses.

By far the most common, predictable and best-studied settings for restraining and releasing bends occur in continental-boundary strike-slip fault systems, where arrays of two to eight en échelon pull-apart basins mark transtensional fault segments and single and sometimes multiple large restraining bends mark transpressional segments; fault areas of transtension versus transpression are determined by the intersection angles between small circles about the interplate pole of rotation and the trend of the strike-slip fault system. These longer and more continuous boundary strike-slip systems also exhibit a widespread pattern of ‘paired bends’ or ‘sidewall ripouts’, or adjacent zones of pull-aparts and restraining bends—that range in along-strike-scale from kilometres to hundreds of kilometres. En échelon arrays of pull-apart basins are also observed on active ‘leaky’ or transtensional oceanic transforms, but restraining bends are rarely observed. In indent-linked strike-slip settings, strike-slip fault traces bounding escaping continental fragments tend to be more arcuate, less-continuous, and more splayed – but paired bends are common. Trench-linked strike-slip fault patterns closely mimic the trends of the subduction zone; these strike-slip faults can vary from long and continuous to short and arcuate, depending on the trace of the adjacent subduction zone. Paired bends are also observed in this setting. Bends on active, cratonic strike-slip fault form isolated, seismically active structures that act as ‘stress concentrators’ for intraplate stress. Cratonic strike-slip faults are generally not associated with pull-apart basins, and therefore paired bends are not observed in this setting.

The most likely geological models for the formation of releasing, restraining bends, and paired bends along boundary and trench-linked strike-slip faults include:

1. progressive linkage of en échelon shears within a young evolving shear zone; this model is not applicable to older strike-slip fault traces that have accumulated significant, lateral fault offsets;

2. formation of lenticular 'sidewall ripout' structures at scales ranging from outcrop to regional; ripouts are thought to form as a response to adherence or sticking along an adjacent and relatively straight strike-slip fault zone; this structural concept may help to explain the large number of paired bends embedded within strike-slip systems, sinu-soidal curvature along the traces of many strike-slip faults, and the episodic nature of lateral shifts in the main strike-slip fault zone;
3. interaction of propagating strike-slip faults with pre-existing crustal structures such as ancient rift basins. Propagation of new strike-slip faults and interaction with older structures may occur on plate boundary, indent-linked, and trench-linked strike-slip faults; and
4. concentration of regional maximum compressive stress on pre-existing, basement fault trends in stable cratonic areas can produce active restraining-bend structures; periodic release of these bend-related stress concentrations is one of the leading causes of intraplate earthquakes within otherwise stable cratons.

The bulges [restraining bends] and gaps [releasing bends] became apparent when the [tectonic] restoration was made. The bulges appeared as areas of discrepancy from which sections of the blocks have been eliminated by faulting, folding, or warping, while the gaps are now occupied by bodies of water, or by sediments and lava. From: A. M. Quennell, 1956, *Tectonics of the Dead Sea Rift, 20th Session of the International Geological Congress*, Mexico City, Mexico.

Significance of strike-slip bends and unresolved questions

Previous work

The description and tectonic origin of releasing and restraining bends on strike-slip faults have fascinated researchers since their first recognition on active strike-slip faults by A. M. Quennell (1956) along the Dead Sea fault in 1956 (cf. Girdler 1989); by Kingma (1958) and Lensen (1958) along the Alpine–Macquarie Ridge fault system of New Zealand; and by Crowell (1974*a*) along the San Andreas fault zone of California (Fig. 1). Kingma (1958) postulated that the formation of basins at fault gaps, and their eventual deformation at fault overlaps, ultimately leads to the difficulty of recognizing and reconstructing ancient strike-slip zones in the rock record. Crowell (1974*a, b*, 1976) applied the early three-dimensional concepts of strike-slip fault deformation of Quennell, Kingma and Lensen, and larger-scale, early plate-tectonic concepts outlined by Carey (1958), to Late Cenozoic tectonics and sedimentation along the San Andreas fault zone in California (Fig. 1*a*). Using examples from the active San Andreas fault system, Crowell (1974*a, b*) named areas of extensional deformation at fault bends 'releasing bends' (Fig. 1*b*) and areas of convergent deformation at fault bends 'restraining bends' (Fig. 1*c*). Later workers, like Cowgill *et al.* (2004*a, b*) also use Crowell's (1974*a*) terms 'double bend' for a bend

developed at an en échelon offset in a fault, and 'single bend' for a fault that terminates within the bend.

Objectives of this paper

This overview paper to the special volume on strike-slip bends synthesizes 50 years of bend-related research, spanning many fields of the geosciences, and focuses attention on important but poorly answered questions about the tectonics of large-offset oceanic and continental strike-slip faults that include:

1. what tectonic processes are responsible for producing prominent or subtle releasing and restraining bends along mainly straight strike-slip faults? (Fig. 1);
2. are these tectonic processes the same or do they differ in the five main tectonic settings for active strike-slip faults (Woodcock 1986; Fig. 2 & Table 1): for example, does a strike-slip fault formed in a collisional setting show the same type of bends as those seen on an oceanic transform fault or a strike-slip fault formed in a fore-arc setting by oblique subduction?
3. why are adjacent restraining and releasing or 'paired bends' produced along the length of many active and ancient strike-slip faults? (Fig. 1); and
4. how do single and paired bends evolve morphologically and structurally with the progressive offset of the strike-slip fault?

Scope of this paper

The central topic of this paper is the tectonic setting, progressive geological evolution, and possible tectonic controls of strike-slip releasing- and restraining bends occurring in natural settings (Fig. 1*a* & Table 2). Many of these descriptions of natural

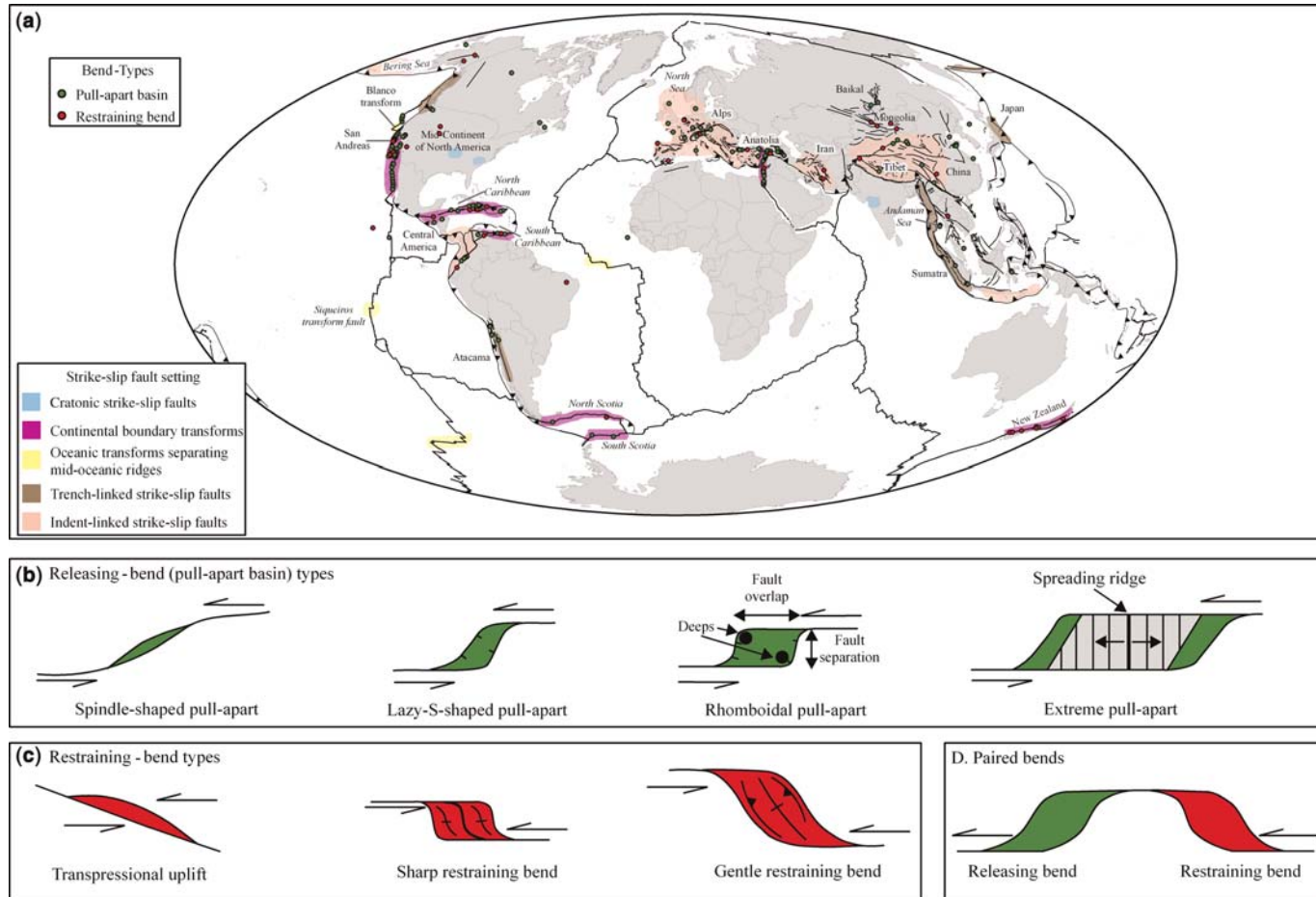


Fig. 1. (a) Global map showing distribution of pull-apart basins (green dots) and restraining bends (red dots) on active and ancient faults: (1) continental boundary transforms in purple; (2) oceanic transforms separating mid-oceanic ridges in yellow; (3) trench-linked strike-slip faults in brown; (4) indent-linked strike-slip faults in pink; and (5) cratonic strike-slip faults in blue. Most of the restraining and releasing bends that are shown on this map are active – although also included are well-studied ancient inactive examples. The main strike-slip fault systems discussed in this paper are labelled. (b) Releasing bend (pull-apart basin) types. Spindle-shaped or nucleating pull-apart, lazy-S-shaped pull-apart, rhomboidal pull-apart and extreme pull-apart. Refer to text for discussion. (c) Restraining-bend types: transpressional uplift, sharp restraining bend, gentle releasing bend. Refer to text for discussion. (d) Paired bend, or releasing and restraining bend. Refer to text for discussion.

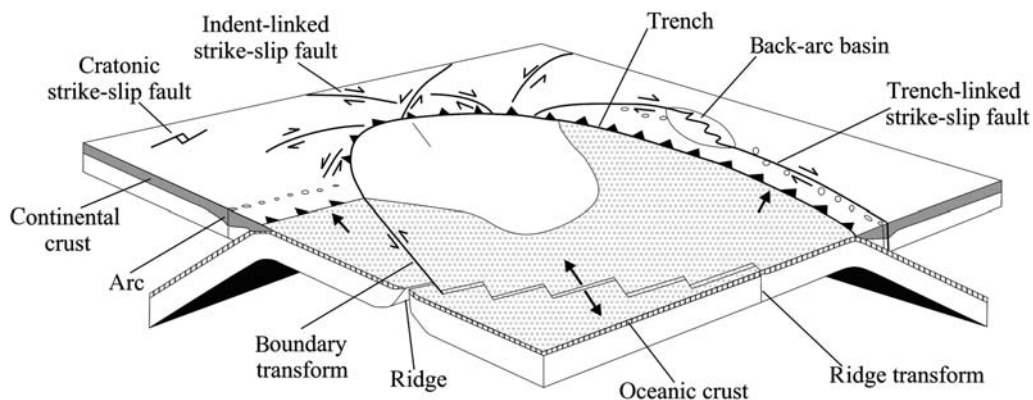


Fig. 2. Strike-slip fault tectonic settings, modified from Woodcock (1986).

bends (e.g. Segall & Pollard 1980; Aydin & Nur 1982; Ben-Avraham & Zoback 1992; Dooley & McClay 1997; McClay & Bonora 2001) have focused on the geology and mechanics of individual and geographically isolated bend structures and do not attempt to relate bends to their tectonic setting, their relationship to regional plate vectors, or their regional geology (Fig. 2). For this reason, this review uses a large number of maps of regional strike-slip fault systems to emphasize the occurrences of releasing and restraining bends that do not occur singly but instead occur in arrays of multiple, en échelon basins or bends. Such en échelon bend arrays indicate that regional tectonic controls and regional plate vectors can be important factors of bend development for along-fault strike distances of tens to hundreds of kilometres (Le Pichon & Francheteau 1978; Garfunkel 1981; Mann *et al.* 1983; Woodcock 1986; Fig. 2).

In the maps and descriptions in this paper, I will emphasize bend structures that have been described from strike-slip systems other than the well-studied San Andreas fault zone of the western USA and Mexico (Fig. 1), which is the most commonly used example of a continental strike-slip system in various chapters on strike-slip faulting in structural and tectonic undergraduate and graduate textbooks: Twiss & Moores (1992); Hatcher (1995); Ingersoll & Busby (1995); Moores & Twiss (1995); Davis & Reynolds (1996); Yeats *et al.* (1997); and Allen & Allen (2005).

The non-US areas of recent bend observations discussed in this paper include the Caribbean, the submarine Macquarie Ridge south of New Zealand, the Scotia Sea, the Tibetan Plateau, the Mongolian Altai, the Lake Baikal region, SE Asia, Sumatra, and the Philippines (Fig. 1 & Table 2). It is especially important to include these non-San Andreas examples in a review of

this type, since the San Andreas fault system represents only one of the five types of strike-slip faults that are described here, and therefore its bends may not be representative for all types of strike-slip fault (Table 1).

Because this paper focuses on strike-slip bends, it is not intended to be a comprehensive review of all aspects of strike-slip faulting, seismology, geodesy, deformation and sedimentation. The interested reader requiring a broader review of strike-slip faulting is referred to the above-mentioned textbooks, along with specialized volumes and review papers on strike-slip faulting by: Ballance & Reading (1980), Sylvester (1984), Lowell (1985), Biddle & Christie-Blick (1985), Harding (1985), Woodcock (1986), Sylvester (1988), Woodcock & Schubert (1989), Harding (1990), Holdsworth *et al.* (1998), and Storti *et al.* (2003). A concise and well-illustrated summary of the morphology, seismicity and active tectonics of many of the world's best-studied strike-slip faults is given in a special issue of *Annals of Tectonics* (1992, edited by Bucknam & Hancock 1992).

Strike-slip fault terminology

Fault terms used in this paper

Any description and discussion of strike-slip bends requires the use of a specialized vocabulary that has evolved over the past 50 years, and classification schemes for strike-slip faults that have been proposed and modified since the late 1980s. For a paper describing so many different strike-slip fault zones, I will adopt a specific set of terms that are illustrated on Figure 1 and used consistently throughout this paper.

Table 1. Characteristics and examples of four major classes of strike-slip faults discussed in this paper (modified from Woodcock 1986) (cf. Figure 2)

	Oceanic transforms separating mid-oceanic spreading ridges	Continental boundary strike-slip faults	Indent-linked strike-slip faults	Trench-linked strike-slip faults	Cratonic strike-slip faults
Crustal type	Ocean–ocean	Continent–continent, continent–ocean	Continent–continent, arc–continent	Continent–continent, arc–continent	Continent–continent
Active duration	Usually <5 Ma	Tens of millions of years; thousands of kilometres	Tens of millions of years; 10–200 km	Tens of millions of years	Tens of millions of years
Strike-slip offset	Typically <100 km but some much larger; later dip-slip	New or reactivated	Reactivate old structures where possible, but some new strands	Hundreds of kilometres	Kilometres
New or reactivated?	New, but may nucleate at old rift margin features	Very variable, non-marine to deep marine clastics or non-deposition	Non-marine clastic and playa sedimentation	New, but including old segments: may nucleate on or localize arc	Reactivated faults commonly inherited from rifts or plutons
Coeval sedimentation in fault zone	Pelagic carbonates and cherts, metaliferous sediments, ophiolitic screes	Rare in transpressive zones, variable but often alkaline basalt in transtension	Rare silicic lavas	Arc-derived clastics, non-marine to marine or non-depositional	Not observed
Coeval volcanism in fault zone	Cut MORB tholeiitic basalts; coeval more alkaline basalts	Not observed	Not observed	Arc tholeiites or calc-alkaline associations; high-K shoshonites may be common	Not observed
Coeval plutonism in fault zone	Ophiolitic plutonics; serpentinite diapirs on faults	Rare in transpression	Probably S-type granites	Gabbros and I-type granites	Not observed
Review papers	DeLong <i>et al.</i> (1979); Bonatti <i>et al.</i> (1979); Fox & Gallo (1984); Garfunkel (1986); Searle (1986); Fornari <i>et al.</i> (1989)	Mann <i>et al.</i> (1983); Sylvester (1988); Yeats <i>et al.</i> (1997)	Sylvester (1988); Yeats <i>et al.</i> (1997)	Fitch (1972); Karig (1979); Jarrard (1986); Yeats <i>et al.</i> (1997)	Marshak <i>et al.</i> (2003); Gangopadhyay & Talwani (2003); Schulte & Mooney (2005)

Table 2. Locations of select basins and bends shown in regional maps

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
Active releasing bends							
1	Spindle-shaped pull-apart	Active	L. Quaternary	Algeciras Basin	Colombia	Algeciras	Velandia <i>et al.</i> (2005)
2	Spindle-shaped pull-apart	Active	L. Quaternary	Ararat Basin	Turkey; Armenia; Iran; Azerbaijan	Sardarapat–Nakhichevan; Dogubayazit; Maku	Karakhianian <i>et al.</i> (2004)
3	Spindle-shaped pull-apart	Active	Quaternary	Balsillas Basin	Colombia	Algeciras	Velandia <i>et al.</i> (2005)
4	Spindle-shaped pull-apart	Active	Neogene	Bichuan	China	Nanding River–Xiaojin	Deng <i>et al.</i> (1986)
5	Spindle-shaped pull-apart	Active	Quaternary	Cantil Valley composite pull-apart basin; Koehn Lake Basin	USA (California)	Garlock	Aydin & Nur (1982); Aydin & Nur (1985)
6	Spindle-shaped pull-apart	Active	Quaternary	Clonard Basin	Haiti	Enriquillo–Plantain Garden	Mann <i>et al.</i> (1983); Mann <i>et al.</i> (1995)
7	Spindle-shaped pull-apart	Active	Miocene–L. Quaternary	Eastern Alps	Austria	Mur–Murz	Sachsenhofer <i>et al.</i> (2003)
8	Spindle-shaped pull-apart	Active	Quaternary	El Paraiso Basin	Colombia	Algeciras	Velandia <i>et al.</i> (2005)
9	Spindle-shaped pull-apart	Active	Quaternary	E’shan Basin	China	Qujiang	Swanson (2005)
10	Spindle-shaped pull-apart	Active	Quaternary	Lago Fagnano Basin	Chile	Magallanes–Fagnano	Lodolo <i>et al.</i> (2003)
11	Spindle-shaped pull-apart	Active	Late Quaternary	Lake Glynn Wye Basin	New Zealand	Hope	Freund (1971); Cowan (1990); Cowan (1994)
12	Spindle-shaped pull-apart	Active	L. Quaternary	Mesquite Basin	USA (California)	Brawley–Imperial	Fuis <i>et al.</i> (1982); Mann <i>et al.</i> (1983); Dooley & McClay (1997)
13	Spindle-shaped pull-apart	Active	L. Quaternary	Neiva Basin	Colombia	Garzon	Chorowicz <i>et al.</i> (1996)
14	Spindle-shaped pull-apart	Active	L. Quaternary	North Aegean Trough	Aegean Sea	Athos	Dooley & McClay (1997)

15	Spindle-shaped pull-apart	Active	L. Quaternary	South Golbaf	Iran	Gowk	Walker & Jackson (2004)
16	Spindle-shaped pull-apart	Active	L. Quaternary	Various unnamed basins	Iceland	Selsund	Angelier <i>et al.</i> (2004)
17	Lazy-S pull-apart	Active	Oligocene–L. Quaternary	Lake Baikal	Russia	Baikal	Petit <i>et al.</i> (1996); Lesne <i>et al.</i> (1998)
18	Lazy-S pull-apart	Active	Pliocene–L. Quaternary	Salina del Fraile	Argentina	El Fraile	Dooley & McClay (1997); Reijs & McClay (2003)
19	Lazy-S pull-apart	Active	Quaternary	San Juan de Villalobos–Yunguillo Basin	Colombia	Yunguillo	Velandia <i>et al.</i> (2005)
20	Lazy-S pull-apart	Active	Neogene	Siabu Basin	Indonesia	Sumatra	Lelgemann <i>et al.</i> (2000)
21	Lazy-S pull-apart	Active	Quaternary	Siqueiros (four basins)	Pacific Ocean	Siqueiros	Fornari <i>et al.</i> (1989)
22	Lazy-S pull-apart	Active	L. Quaternary	Unnamed basins (South Scotia Ridge)	Antarctica	South Scotia Ridge	Galindo-Zaldivar <i>et al.</i> (1996); Galindo-Zaldivar <i>et al.</i> (2002)
23	Lazy-S pull-apart	Active	Eocene–L. Quaternary	Virgin Islands Basin	Virgin Islands	Unnamed	Jany <i>et al.</i> (1990)
24	Lazy-Z pull-apart	Active	Oligocene–L. Quaternary	Bohai Basin	China	Tanlu	Allen <i>et al.</i> (1997)
25	Lazy-Z pull-apart	Active	Neogene–L. Quaternary	Central Basin	Turkey	North Anatolian	Barka & Gulen (1989); Armijo <i>et al.</i> (1989); Le Pichon <i>et al.</i> (2001); Armijo <i>et al.</i> (2002); Rangin <i>et al.</i> (2004)
26	Lazy-Z pull-apart	Active	Cenozoic	Cerro de la Mica	Chile	Atacama	McClay & Bonora (2001)
27	Lazy-Z pull-apart	Active	Quaternary	Clear Lake	USA (California)	Mountain	McLaughlin & Nilsen (1982); Hearne <i>et al.</i> (1988)
28	Lazy-Z pull-apart	Active	L. Plio–Quaternary	Dagg Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2001); Barnes <i>et al.</i> (2005)
29	Lazy-Z pull-apart	Active	Miocene–Quaternary	Death Valley Basin	USA (California)	Southern Death Valley; Fish Lake Valley–Northern Death Valley–Furnace Creek	Burchfiel Stewart (1966); Christie-Blick & Biddle (1985)
30	Lazy-Z pull-apart	Active	L. Plio–Quaternary	Five Fingers Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
31	Lazy-Z pull-apart	Active	L. Plio–Quaternary	George Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2005)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
32	Lazy-Z pull-apart	Active	L. Oligocene to M.'Miocene	Gulf of Paria Basin	Venezuela	El Pilar; Central Range	Babb & Mann (1999); Flinch <i>et al.</i> (1999)
33	Lazy-Z pull-apart	Active	L. Quaternary	Karakoram Basin	China	Karakoram	Murphy & Burgess (2006)
34	Lazy-Z pull-apart	Active	Pliocene (?)–Quaternary	La Gonzalez Basin	Venezuela	Bocono	Christie-Blick and Biddle (1985)
35	Lazy-Z pull-apart	Active	L. Quaternary	Lake Tahoe	USA (California)	Walker Lane	Unruh <i>et al.</i> (2003)
36	Lazy-Z pull-apart	Active	L. Plio–Quaternary	Looking Glass Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
37	Lazy-Z pull-apart	Active	Neogene–L. Quaternary	Mor	Israel	Dead Sea	Bartov & Sagy (2004)
38	Lazy-Z pull-apart	Active	Late Quaternary	Mount Lyford duplex	New Zealand	Hope	Eusden <i>et al.</i> (2000)
39	Lazy-Z pull-apart	Active	Early Pliocene–L. Quaternary	Niksar Basin	Turkey	North Anatolian	Mann <i>et al.</i> (1983); Tatar <i>et al.</i> (1995); Barka <i>et al.</i> (2000); Kocyigit & Erol (2001)
40	Lazy-Z pull-apart	Active	L. Quaternary	Offshore Santa Catalina Island	USA (California)	San Diego Trough–Catalina	Legg <i>et al.</i> (2007)
41	Lazy-Z pull-apart	Active	Neogene–L. Quaternary	Panamint	USA (California)	Walker Lane	Burchfiel <i>et al.</i> (1987)
42	Lazy-Z pull-apart	Active	L. Miocene–L. Quaternary	Ridge Basin	USA	San Andreas; San Gabriel	May <i>et al.</i> (1993)
43	Lazy-Z pull-apart	Active	L. Quaternary	Round Valley	USA (California)	Bartlett Springs	McLaughlin & Nilsen (1982); Hearne <i>et al.</i> (1988)
44	Lazy-Z pull-apart	Active	Neogene–L. Quaternary	Saline valleys	USA (California)	Walker Lane	Burchfiel <i>et al.</i> (1987)
45	Lazy-Z pull-apart	Active	L. Plio–Quaternary	Secretary Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
46	Lazy-Z pull-apart	Active	Quaternary	Sunda Strait	Sumatra	Sumatran	Leigemann <i>et al.</i> (2000)
47	Lazy-Z pull-apart	Active	Oligocene–L. Quaternary	Tunka	Russia	Tunka	Larroque <i>et al.</i> (2001)
48	Rhomboidal pull-apart	Active	Quaternary	Alarcon Basin (Gulf of California)	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
49	Rhomboidal pull-apart	Active	L. Quaternary	Altamira–Pitalito Basin	Colombia	Granadillo; El Cedro; Pitalito	Velandia <i>et al.</i> (2005)

50	Rhomboidal pull-apart	Active	Oligocene–L. Quaternary	Andamen Sea	Thailand	Seuliman; Sumatra	Curray (2005)
51	Rhomboidal pull-apart	Active	L. Quaternary	Aragonese Basin	Gulf of Aqaba	Arava	Ben-Avraham (1985); Erhardt <i>et al.</i> (2005); Garfunkel (2005)
52	Rhomboidal pull-apart	Active	E. Miocene–L. Quaternary	Arnona Basin	Gulf of Aqaba	Arava	Ben-Avraham (1985)
53	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Baitiquiri	Caribbean Sea	Oriente	Calais & Mercier de Lepinay (1991)
54	Rhomboidal pull-apart	Active	Quaternary	Barahta	Israel	Rachaya; Si'on; Shab'a	Heimann <i>et al.</i> (1990)
55	Rhomboidal pull-apart	Active	Not given in paper	Bayburt Basin	Turkey	Kelkit–Coruh	Kocyigit & Erol (2001)
56	Rhomboidal pull-apart	Active	Quaternary	Bend region	USA (California)	San Clemente	Legg <i>et al.</i> (2007)
57	Rhomboidal pull-apart	Active	L. Quaternary	Beng Co	China		Armijo <i>et al.</i> (1989)
58	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Cabo Cruz	Caribbean Sea	Oriente	Calais & Mercier de Lepinay (1991)
59	Rhomboidal pull-apart	Active	L. Oligocene or E. Miocene to L. Quaternary	Cariaco Basin	Venezuela	Moron; El Pilar; San Sebastian	Schubert (1982); Mann <i>et al.</i> (1983); Jaimes (2003)
60	Rhomboidal pull-apart	Active	Not given in paper	Carmen Basin (Gulf of California)	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
61	Rhomboidal pull-apart	Active	Pliocene	Cascadia	Offshore USA (Oregon)	Blanco transform	Embley & Wilson (1992)
62	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Chivirico	Caribbean Sea	Oriente	Calais Mercier de Lepinay (1991)
63	Rhomboidal pull-apart	Active	L. Quaternary	Cholame Valley Basin	USA (California)	San Andreas	R. Brown & Vedder (1967); Shedlock <i>et al.</i> (1990); Graymer <i>et al.</i> (2007)
64	Rhomboidal pull-apart	Active	Miocene–Pliocene	Cinarcik Basin	Turkey	North Anatolian	Okay <i>et al.</i> (2000); Le Pichon <i>et al.</i> (2001); Seeber <i>et al.</i> (2006)
65	Rhomboidal pull-apart	Active	Neogene	Coyote Ridge	USA (California)	Coyote Ridge (San Jacinto)	Sharp (1975)
66	Rhomboidal pull-apart	Active	Neogene–L. Quaternary?	Daisy Bank	USA	Daisy Bank	Goldfinger <i>et al.</i> (1996)
67	Rhomboidal pull-apart	Active	E. Miocene–L. Quaternary	Dakar Basin	Gulf of Aqaba	Dead Sea	Ben-Avraham (1985)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
68	Rhomboidal pull-apart	Active	Late Miocene–L. Quaternary	Damxung Corridor Basin	China	Damxung–Jiali	Armijo <i>et al.</i> (1986); Edwards & Ratschbacher (2005)
69	Rhomboidal pull-apart	Active	E. Miocene–E. Pliocene (L. Quaternary)	Dead Sea Basin	Jordan	Dead Sea; Jordan; Arava	Ten-Brink & Ben-Avraham (1989); Lazar <i>et al.</i> (2006)
70	Rhomboidal pull-apart	Active	Not given in paper	Delfin Basin (Gulf of California)	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
71	Rhomboidal pull-apart	Active	Tertiary	Dungun Basin	Malaysia	Dungun	Dooley & McClay (1997)
72	Rhomboidal pull-apart	Active	Quaternary	East Blanco	Offshore USA (Oregon)	Blanco transform	Embley & Wilson (1992)
73	Rhomboidal pull-apart	Active	Pliocene–Quaternary	Erciyes Basin	Turkey	Central Anatolian	Kocyigit & Erol (2001)
74	Rhomboidal pull-apart	Active	Quaternary	Erzincan Basin	Turkey	Tercan–Askale	Aydin & Nur (1982); Barka & Gulen (1989); Kocyigit & Erol (2001)
75	Rhomboidal pull-apart	Active	Not given in paper	Farallon Basin (Gulf of California)	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
76	Rhomboidal pull-apart	Active	Not given in paper	Ghab Basin	Syria	Dead Sea	Brew <i>et al.</i> (2001); Adiyaman & Chorowicz (2002); Westaway (2004)
77	Rhomboidal pull-apart	Active	L. Quaternary	Golden Gate	USA (California)	San Andreas	Zoback <i>et al.</i> (1999); Wakabayashi (2007)
78	Rhomboidal pull-apart	Active	Quaternary	Gorda	Offshore USA (Oregon)	Blanco transform	Embley & Wilson (1992)
79	Rhomboidal pull-apart	Active	Not given in paper	Guaymas Basin (Gulf of California)	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
80	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Gulu	China	Gulu	Edwards & Ratschbacher (2005)
81	Rhomboidal pull-apart	Active	Quaternary	Hamner Basin	New Zealand	Hope	Freund (1971); Wood <i>et al.</i> (1994)
82	Rhomboidal pull-apart	Active	L. Quaternary	Homestead Valley Basin	USA (California)	Homestead Valley; Johnson Valley	Sowers <i>et al.</i> (1994)
83	Rhomboidal pull-apart	Active	Not given in paper	Hula Basin	Israel; Lebanon	Dead Sea	Heimann & Ron (1987); Zilberman <i>et al.</i> (2000)

84	Rhomboidal pull-apart	Active	Neogene-L. Quaternary	Jianchuan–Hequin	China	Xiaojinhe	Deng <i>et al.</i> (1986)
85	Rhomboidal pull-apart	Active	Not given in paper	Karasu Basin	Turkey	Tercan–Askale	Kocyigit & Erol (2001)
86	Rhomboidal pull-apart	Active	Not given in paper	Karlioiva Basin	Turkey	East Anatolian	Kocyigit & Erol (2001)
87	Rhomboidal pull-apart	Active	Not given in paper	Kinneret Zemah Basin	Israel	Dead Sea	Rotstein & Bartov (1989)
88	Rhomboidal pull-apart	Active	Not given in paper	Lake Elsinore Basin	USA (California)	Elsinore	Mann <i>et al.</i> (1983)
89	Rhomboidal pull-apart	Active	Not given in paper	Lake Hazar Basin	Turkey	East Anatolian	Mann <i>et al.</i> (1983); Hempton & Dunne (1983)
90	Rhomboidal pull-apart	Active	L. Miocene	Lake Matano Basin	Indonesia	Matano	Parkinson & Dooley (1996)
91	Rhomboidal pull-apart	Active	L. Miocene	Lake Poso Basin	Indonesia	Poso	Parkinson & Dooley (1996)
92	Rhomboidal pull-apart	Active	Quaternary	Laohuyiaoxian	China	Xiaokou	Zhang <i>et al.</i> (1989)
93	Rhomboidal pull-apart	Active	L. Quaternary	Little Lake Valley	USA (California)	Maacama	McLaughlin & Nilsen (1982); Hearne <i>et al.</i> (1988)
94	Rhomboidal pull-apart	Active	L. Quaternary	Little Sulfur Creek	USA (California)	Maacama	McLaughlin & Nilsen (1982); Hearne <i>et al.</i> (1988)
95	Rhomboidal pull-apart	Active	Miocene–L. Quaternary	Macquarie Ridge	Australia	Australian–Antarctic plate boundary	Massell <i>et al.</i> (2000); Daczko <i>et al.</i> (2003)
96	Rhomboidal pull-apart	Active	Not given in paper	Malatya Basin	Turkey	Malatya–Ovacik	Kocyigit & Erol (2001)
97	Rhomboidal pull-apart	Active	Not given in paper	Mazatlan Basin	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
98	Rhomboidal pull-apart	Active	Pliocene–L. Quaternary	Merced	USA (California)	San Andreas	Wakabayashi (2007)
99	Rhomboidal pull-apart	Active	Quaternary	Mirogoane Lakes Basin	Haiti	Enriquillo–Plantain Garden	Mann <i>et al.</i> (1983); Mann <i>et al.</i> (1995)
100	Rhomboidal pull-apart	Active	L. Plio–Quaternary	Nancy Basin	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
101	Rhomboidal pull-apart	Active	Quaternary	Northern Gulf of Aqaba	Israel	Dead Sea	Ehrhardt <i>et al.</i> (2005)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
102	Rhomboidal pull-apart	Active	Quaternary	Not given in paper	New Zealand	Alpine; Boo Boo; Marlborough	Barnes (2005)
103	Rhomboidal pull-apart	Active	L. Quaternary	Olema	USA (California)	San Andreas	Wakabayashi (2007)
104	Rhomboidal pull-apart	Active	Not given in paper	Pescadero	USA (California)	San Andreas	Mann <i>et al.</i> (1983)
105	Rhomboidal pull-apart	Active	Quaternary	Pitalito Basin	Colombia	San Francisco	Velandia <i>et al.</i> (2005)
106	Rhomboidal pull-apart	Active	Quaternary–L. Quaternary	Rio Lempa Basin	El Salvador	El Salvador (San Vicente segment and Berlin segment)	Corti <i>et al.</i> (2005)
107	Rhomboidal pull-apart	Active	L. Quaternary	San Jacinto Basin	USA (California)	Casa Loma; Claremont	Langenheim <i>et al.</i> (2004)
108	Rhomboidal pull-apart	Active	Late Quaternary	San Pablo Bay	USA (California)	Hayward; Rodgers Creek	Parsons <i>et al.</i> (2003)
109	Rhomboidal pull-apart	Active	L. Quaternary	San Pedro Martir (Gulf of California)	Mexico	San Pedro	Mann <i>et al.</i> (1983); Legg <i>et al.</i> (2007)
110	Rhomboidal pull-apart	Active	Pliocene–Quaternary	Sea of Galilee	Israel	Dead Sea	Hurwitz <i>et al.</i> (2002)
111	Rhomboidal pull-apart	Active	Quaternary	Sibundoy Basin	Colombia	Sibundoy; San Francisco	Velandia <i>et al.</i> (2005)
112	Rhomboidal pull-apart	Active	Quaternary	Singkarak	Indonesia	Sumatran; Metawai	Sieh & Natawidjaja (2000)
113	Rhomboidal pull-apart	Active	Eocene–L. Quaternary	St Croix Basin	St Croix	Right-lateral strike-slip	Jany <i>et al.</i> (1990)
114	Rhomboidal pull-apart	Active	Quaternary	Surveyor Basin	Offshore USA (Oregon)	Blanco transform	Embley & Wilson (1992)
115	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Swan Islands	Honduras	Swan Islands	Mann <i>et al.</i> (1991)
116	Rhomboidal pull-apart	Active	L. Quaternary	Tabriz Basin	Iran	North Tabriz	Karakhanian <i>et al.</i> (2004)
117	Rhomboidal pull-apart	Active	Pliocene–Early Quaternary	Tasova–Erbaa	Turkey	North Anatolian	Barka <i>et al.</i> (2000)

118	Rhomboidal pull-apart	Active	Miocene-Pliocene	Tekirdag Basin	Turkey	North Anatolian	Okay <i>et al.</i> (1999); Le Pichon <i>et al.</i> (2001); Seeber <i>et al.</i> (2006)
119	Rhomboidal pull-apart	Active	L. Oligocene to E. Miocene–L. Quaternary	Tesac Basin	Atlantic offshore	Magallanes–Fagnano	Lodolo <i>et al.</i> (2003)
120	Rhomboidal pull-apart	Active	Quaternary	Tiburon river valley	Haiti	Enriquillo–Plantain Garden	Mann <i>et al.</i> (1983); Mann <i>et al.</i> (1995)
121	Rhomboidal pull-apart	Active	E. Miocene–L. Quaternary	Tiran Basin	Gulf of Aqaba	Dead Sea	Ben-Avraham (1985)
122	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Unnamed	China	Xianshuihe; Ganzi–Yushu	Allen <i>et al.</i> (1991)
123	Rhomboidal pull-apart	Active	Quaternary–L. Quaternary	Vanadzor	Armenia	Pambak-Sevan–Sunik	Karakhanian <i>et al.</i> (2004)
124	Rhomboidal pull-apart	Active?	E. Quaternary–L. Quaternary?	Various unnamed pull-apart basins	USA (Nevada)	Olinghouse	Aydin & Nur (1982)
125	Rhomboidal pull-apart	Active	Not given in paper	Wagner (Gulf of California)	Mexico	San Andreas	Mann <i>et al.</i> (1983)
126	Rhomboidal pull-apart	Active	E. Quaternary–L. Quaternary	Walker Lake Basin	USA (Nevada)	Walker Lane	Link <i>et al.</i> (1985)
127	Rhomboidal pull-apart	Active	Quaternary	West Blanco Basin	Offshore USA (Oregon)	Blanco transform	Embley & Wilson (1992)
128	Rhomboidal pull-apart	Active	Late Neogene–L. Quaternary	West Jamaica (Hendrix)	Jamaica	Duanvale; Swan Islands	Mann <i>et al.</i> (2007a)
129	Rhomboidal pull-apart	Active	Neogene–L. Quaternary	Windward Passage	Caribbean Sea	Oriente	Calais & Mercier de Lepinay (1991)
130	Spindle-shaped pull-apart	Active	Late Quaternary	Poplars basin	New Zealand	Hope	Cowan (1990); Cowan (1994)
131	Extreme pull-apart	Active	Neogene–L. Quaternary	Andaman Sea	India; Thailand	Sumatra–Sagaing	Curry (2005)
132	Extreme rhomboidal pull-apart	Active	Eocene–L. Quaternary	Cayman Trough	Northern Caribbean Sea	Oriente; Swan	Rosencrantz <i>et al.</i> (1988); Leroy <i>et al.</i> (2000)
Ancient releasing bends							
133	Spindle-shaped pull-apart	Ancient	Eocene	Chiwaukee Basin	USA (Washington)	Entiat; Leavenworth	Johnson (1985)
134	Spindle-shaped pull-apart	Ancient	Eocene	Chuckanut–Puget–Naches Basin	USA (Washington)	Puget; Straight Creek	Johnson (1985)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
135	Spindle-shaped pull-apart	Ancient	L. Palaeozoic–Quaternary	Danish Basin	Sweden; Bornholm; Baltic Sea	Colonus Shale Trough; Romeneasen; Kullen Rjngsjon Andrawum; Christianso	Erlstrom <i>et al.</i> (1997)
136	Spindle-shaped pull-apart	Ancient	?–Quaternary	Dongchuan Basin	China	East Xiaojian	Swanson (2005)
137	Spindle-shaped pull-apart	Ancient	Ordovician–Devonian	Gaspe Basin	Canada	Lac Cascapedia; Mont Albert; Shickshock Sud	Sacks <i>et al.</i> (2004)
138	Lazy-S pull-apart	Ancient	Mesozoic	Caleta Coloso Basin	Chile	Bolfin; Jorgillo; Coloso	Cembrano <i>et al.</i> (2005)
139	Lazy-S pull-apart	Ancient	Oligocene–E. Miocene	Salar Grande Basin	Chile	Atacama	Reijs & McClay (1998)
140	Lazy-S pull-apart	Ancient	Archaean	Whim Creek Basin	Australia	Loudens	Krapez & Eisenlohr (1998)
141	Lazy-Z pull-apart	Ancient	Jurassic–E. Cretaceous	Chaoshui Basin	China	North Qilian Shan	Vincent & Allen (1999)
142	Lazy-Z pull-apart	Ancient	Jurassic–E. Cretaceous	Huahai-Jinta Basin	China	North Qilian Shan	Vincent & Allen (1999)
143	Lazy-Z pull-apart	Ancient	Cretaceous	Jiaolai Basin	China	Tan–Lu; Jimo	Zhang <i>et al.</i> (2003)
144	Lazy-Z pull-apart	Ancient	Jurassic–E. Cretaceous	Minle Basin	China	North Qilian Shan	Vincent Allen (1999)
145	Lazy-Z pull-apart	Ancient	Cretaceous	Zhucheng Basin	China	Tan–Lu; Baichihe; Wulian	Zhang <i>et al.</i> (2003)
146	Rhomboidal pull-apart	Ancient	Miocene–Pliocene	Bir Zreir graben	Egypt	Southern Diagonal; Northern Diagonal	Eyal <i>et al.</i> (1986)
147	Rhomboidal pull-apart	Ancient	Cenozoic	Bovey Basin	England	Stickepath–Lustleigh	Holloway & Chadwick (1986)
148	Rhomboidal pull-apart	Ancient	Late Cretaceous–Cenozoic	Catacamas	Honduras	Guayape	Gordon & Muehlberger (1994)
149	Rhomboidal pull-apart	Ancient	Permian	Collio Basin	Italy	Valcamonica; Giudicarie	Bertoluzza & Perotti (1997)

150	Rhomboidal pull-apart	Ancient	Mesozoic–Cenozoic	Coso Basin	USA (Nevada)	Airport Lake; Owens Valley	Feng & Lees (1998); Unruh <i>et al.</i> (2003);
151	Rhomboidal pull-apart	Ancient	Ordovician–Devonian	Culm Basin	France	Nort-sur-Erdre	Shelley & Bossiere (2001)
152	Rhomboidal pull-apart	Ancient	Quaternary	Dayinshui	China	Haiyuan	Zhang <i>et al.</i> (1989)
153	Rhomboidal pull-apart	Ancient	Triassic	Donaustauf Basin	Germany	Pfahl; Danube	Mattern (2001)
154	Rhomboidal pull-apart	Ancient	Cretaceous	Eumsung Basin	Korea	Kongiu	Ryang & Chough (1999)
155	Rhomboidal pull-apart	Ancient	Jurassic–Cretaceous	Guinea marginal plateau	Guinea	Unnamed	Benkhelil <i>et al.</i> (1995)
156	Rhomboidal pull-apart	Ancient	Eocene	Icotea	Venezuela	Icotea	Escalona & Mann (2003)
157	Rhomboidal pull-apart	Ancient	Jurassic–Cretaceous	Inner Moray Firth Basin	North Sea	Great Glen	Dooley <i>et al.</i> (1999)
158	Rhomboidal pull-apart	Ancient	Jurassic	Kuzigongsu pull-apart	Kazakostan; Kyrgyzstan; China	Karatau/Talas–Fergana	Allen <i>et al.</i> (2001)
159	Rhomboidal pull-apart	Ancient	Carboniferous–Miocene (?)	Magdalena Basin	Canada	Long Range; Belleisle; Kennebecasis; Caldonia	Mann <i>et al.</i> (1983)
160	Rhomboidal pull-apart	Ancient	Triassic–E. Cretaceous	Mazhan Basin	China	Tancheng–Lujian (Tan–Lu)	Hong & Miyata (1999)
161	Rhomboidal pull-apart	Ancient	E. Proterozoic	Nonacho Basin	Canada	Noman; King; Macinnis; Salkeld	Aspler & Donaldson (1985)
162	Rhomboidal pull-apart	Ancient	Palaeogene	Norton Basin	USA (Alaska)	Kaltag	Fisher <i>et al.</i> (1979)
163	Rhomboidal pull-apart	Ancient	Oligocene–E. Miocene	Oschiri Basin	Italy	Olbia	Funedda & Oggiano (2005)
164	Rhomboidal pull-apart	Ancient	Pre-Tertiary	Overton Arm Basin	USA	Lake Mead	Aydin <i>et al.</i> (1990)
165	Rhomboidal pull-apart	Ancient	Tertiary	Petrockstow Basin	England	Stickepath–Lustleigh	Holloway & Chadwick (1986)
166	Rhomboidal pull-apart	Ancient	Not given in paper	Salt Lake Basin	China	Haiyuan	Zhang <i>et al.</i> (1989)
167	Rhomboidal pull-apart	Ancient	Quaternary	Shaoshui Basin	China	Haiyuan	Zhang <i>et al.</i> (1989)
168	Rhomboidal pull-apart	Ancient	L. Jurassic–E. Cretaceous	Soria strike-slip Basin	Spain	Unnamed	Guiraud & Seguret (1985)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
169	Rhomboidal pull-apart	Ancient	Carboniferous–Permian	St Etienne Basin	France	France Massif Central	Mattauer & Matte (1998)
170	Rhomboidal pull-apart	Ancient	Palaeogene	St George	USA (Alaska)	Beringian	Worrall (1991)
171	Rhomboidal pull-apart	Ancient	Late Palaeozoic	Stellarton Basin	Canada	Late Palaeozoic	Waldron (2004)
172	Rhomboidal pull-apart	Ancient	Plio–Quaternary	Upper Rhine Graben	Germany	Left-lateral strike-slip	Schumacher (2002)
173	Rhomboidal pull-apart	Ancient	Tertiary (Miocene)	Vienna Basin	Austria; Czech Republic; Slovakia	Waschberg	Mann <i>et al.</i> (1983); Christie-Blick & Biddle (1985); Royden (1985); Hinsch <i>et al.</i> (2005)
174	Rhomboidal pull-apart	Ancient	E. Miocene–M. Miocene	Virginia City	USA (Nevada)	Comstock; Occidental	Berger (2007)
175	Extreme pull-apart	Ancient	Neogene	Sea of Japan	Japan	Unnamed	Lallemant & Jolivet (1986); Burgmann <i>et al.</i> (1994)
Active restraining bends							
1	Transpressional uplift	Active	Neogene–L. Quaternary	Betic Range	Spain	Trans-Alboran	Keller <i>et al.</i> (1995)
2	Transpressional uplift	Active	Cenozoic–Late Neogene	Border Ranges	USA (Alaska)		Little (1990); Roeske <i>et al.</i> (2003)
3	Transpressional uplift	Active	Late Quaternary	Charwell ‘fault wedge’	New Zealand	Hope	Eusden <i>et al.</i> (2005)
4	Transpressional uplift	Active	Quaternary	Clipperton	Pacific Ocean	Clipperton	Pockalny (1997)
5	Transpressional uplift	Active	Late Neogene–L. Quaternary	Mecca Hills	USA (California)	San Andreas	Sylvester & Smith (1976)
6	Transpressional uplift	Active	Late Neogene–L. Quaternary	Southern Alps	New Zealand	Central Alpine	Little <i>et al.</i> (2005)
7	Sharp restraining bend	Active	Neogene–L. Quaternary	Almacik Mountains	Turkey	North Anatolian	Saribudak <i>et al.</i> (1990)
8	Sharp restraining bend	Active	Neogene–L. Quaternary	Araya–Paria	Venezuela	El Pilar	Audemard <i>et al.</i> (2006)

9	Sharp restraining bend	Active	Oligocene–L. Quaternary	Chainat Ridge	Thailand	Mae Ping	Smith <i>et al.</i> (2007)
10	Sharp restraining bend	Active	L. Quaternary	Middleton Place Summerville	USA (South Carolina)	Woodstock; Ashley River	Gangopadhyay & Talwani (2003); Marshak <i>et al.</i> (2003)
11	Sharp restraining bend	Active	?–L. Quaternary	Ocotillo Badlands	USA (California)	Coyote Creek	Sharp & Clark (1972); Brown & Sibson (1989); Brown <i>et al.</i> (1991); Lutz <i>et al.</i> (2006)
12	Sharp restraining bend	Active	L. Quaternary	Rokko Range	Japan	Unnamed	Sangawa (1986)
13	Sharp restraining bend	Active	Late Neogene–L. Quaternary	Unnamed	Mexico	Polo chic; unnamed	Guzmán-Speziale (2001)
14	Sharp restraining bend	Active	Late Neogene	Villa Vasquez	Hispaniola	Septentrional	Mann <i>et al.</i> (1999)
15	Gentle restraining bend	Active	Pliocene (?)–L. Quaternary	Akato Tagh	China	Altyn Tagh	Cowgill <i>et al.</i> (2004a); Cowgill <i>et al.</i> (2004b)
16	Gentle restraining bend	Active	Pliocene (?)–L. Quaternary	Alpine Fault ‘Big Bend’	New Zealand	Alpine	Campbell (1992); Cowan (1994)
17	Gentle restraining bend	Active	Cenozoic–L. Quaternary	Atacama	Chile	Atacama	Swanson (2005)
18	Gentle restraining bend	Active	Neogene–L. Quaternary	Bingol	Turkey	East Anatolian	Hempton & Dunne (1983); Saroglu <i>et al.</i> (1992)
19	Gentle restraining bend	Active	post-M. Miocene–L. Quaternary	Blue Mountains	Jamaica	Plantain Garden; Blue Mountain; Yallahs; Wagwater	Mann <i>et al.</i> (1985); Mann & Gordon (1996); Mann <i>et al.</i> (2007a)
20	Gentle restraining bend	Active	Quaternary	Chicibu–Sanbagawa belts	Japan	Median; Butsuzo	Mann & Gordon (1996); Yeats <i>et al.</i> (1997)
21	Gentle restraining bend	Active?	Quaternary	Crespi Knoll	USA (California)	Palos Verdes	Legg <i>et al.</i> (2007)
22	Gentle restraining bend	Active	Quaternary	Dagg Ridge	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
23	Gentle restraining bend	Active	Neogene–L. Quaternary	Denali	USA (Alaska)	Denali	Fitzgerald <i>et al.</i> (1993)
24	Gentle restraining bend	Active	Cenozoic–L. Quaternary	Domeyko	Chile	Domeyko	Swanson (2005)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
25	Gentle restraining bend	Active	L. Quaternary	Dumbe	USA	Santa Monica–Dume	Sorlien <i>et al.</i> (2006)
26	Gentle restraining bend	Active	Pliocene–Quaternary	East Bay Hills	USA (California)	Hayward	Aydin & Page (1984)
27	Gentle restraining bend	Active	Not given in paper	Echo Hills	USA (Nevada)	Bitter Spring Valley	McClay & Bonora (2001)
28	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Ecuadorian Andes	Ecuador; Colombia	Pallatanga; Rio Chingual–La Sofia	Ego <i>et al.</i> (1996)
29	Gentle restraining bend	Active	L. Quaternary	Ganos	Turkey	North Antolian	Armijo <i>et al.</i> (1999); Seeber <i>et al.</i> (2004)
30	Gentle restraining bend	Active	Quaternary	George Ridge	New Zealand	Alpine	Barnes <i>et al.</i> (2005)
31	Gentle restraining bend	Active	L. Quaternary	Guarapiche	Venezuela	El Pilar	Audemard (2006)
32	Gentle restraining bend	Active	Cenozoic	Gurvan Bogd mountains	Mongolia	Gurvan Bogd	Cunningham <i>et al.</i> (1996a); Bayasgalan <i>et al.</i> (1999)
33	Gentle restraining bend	Active	Late Quaternary	Hayward–Calaveras	USA (California)	Hayward–Calaveras	Graymer <i>et al.</i> (2007)
34	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Hispaniola	Hispaniola	Septentrional	Mann <i>et al.</i> (1984); Mann <i>et al.</i> (1999); Prentice <i>et al.</i> (2003)
35	Gentle restraining bend	Active	L. Quaternary	Kaikoura Ranges	New Zealand	Hope; Jordan; Kekerengu	Van Dissen & Yeats (1991)
36	Gentle restraining bend	Active	L. Quaternary	Khleng Lan	Thailand	Inthanon; Sukhothai	Smith <i>et al.</i> (2007)
37	Gentle restraining bend	Active	L. Quaternary	Lasuen Knoll	USA (California)	Palos Verdes	Fisher <i>et al.</i> (2004); Legg <i>et al.</i> (2007)
38	Gentle restraining bend	Active	L. Quaternary	Lebanon and Anti-Lebanon ranges	Lebanon; Syria	Dead Sea; Levant; Yammounah	Gomez <i>et al.</i> (2007)
39	Gentle restraining bend	Active	Quaternary	Lome Prieta	USA (California)	San Andreas; Sargent–Berrocal	Schwartz <i>et al.</i> (1990)
40	Gentle restraining bend	Active	Neogene–L. Quaternary	Massif du Macaya	Haiti	Enriquillo–Plantain Garden	Mann <i>et al.</i> (1983); Mann <i>et al.</i> (1995)
41	Gentle restraining bend	Active	L. Quaternary	Mission Hills	USA (California)	Hayward–Calaveras	Andrews <i>et al.</i> (1993); Burgmann <i>et al.</i> (2006); Manaker <i>et al.</i> (2005)

42	Gentle restraining bend	Active	L. Quaternary	Mt Oxford	New Zealand	Porters Pass–Amberley	Cowan (1992); Cowan <i>et al.</i> (1996)
43	Gentle restraining bend	Active	Pliocene–L. Quaternary	Mt. Diablo	USA (California)	Greenville; Concord	Burgmann <i>et al.</i> (2006); Wakabayashi (2007)
44	Gentle restraining bend	Active	Cenozoic	Nemegt Uul	Mongolia	Gobi–Tien Shan	Cunningham <i>et al.</i> (1995); Owen <i>et al.</i> (1999)
45	Gentle restraining bend	Active	L. Quaternary	Palos Verdes Hills	USA (California)	Palos Verdes	Ward (1994); Fisher <i>et al.</i> (2004); Legg <i>et al.</i> (2007)
46	Gentle restraining bend	Active	Miocene	Pangong Range	India	Karakoram	Searle & Phillips (2005)
47	Gentle restraining bend	Active	Tertiary	Pijnacker	Netherlands	Not given in paper	McClay & Bonora (2001)
48	Gentle restraining bend	Active	Late Quaternary	Porters Pass	New Zealand	Porters Pass	Paterson Campbell (1998)
49	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Santa Catalina Island	USA (California)	Santa Catalina	Ward (1994); Mann & Gordon (1996); Fisher <i>et al.</i> (2004); Legg <i>et al.</i> (2007)
50	Gentle restraining bend	Active	Neogene–L. Quaternary	Santa Cruz Mountains	USA (California)	San Andreas	Anderson (1990); Schwartz <i>et al.</i> (1990); Burgmann <i>et al.</i> (1994); Anderson (1994)
51	Gentle restraining bend	Active	L. Cretaceous–L. Cenozoic	Sierra de las Minas	Guatemala	Motagua; Polochic	Aydin & Nur (1982); Mann & Gordon (1996)
52	Gentle restraining bend	Active	Late Neogene–L. Quaternary	South Georgia Island	Antarctica	North Scotia Ridge	Mann <i>et al.</i> (2003)
53	Gentle restraining bend	Active	L. Cenozoic	Sutai Range	Mongolia	Tonhil	Cunningham (1996b); Cunningham <i>et al.</i> (2003c)
54	Gentle restraining bend	Active	Quaternary	Tararua Ranges	New Zealand	Wellington	Beanland (1995)
55	Gentle restraining bend	Active	Eocene–L. Quaternary	Tortola ridge	Caribbean Sea	Anegada passage	Jany <i>et al.</i> (1990)
56	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Transverse Ranges (San Bernardino Mountains)	USA (California)	San Andreas	Spotila & Anderson (2004)
57	Gentle restraining bend	Active	Miocene	Transverse Ranges (San Gabriel Mountains)	USA (California)	San Andreas	Blythe <i>et al.</i> (2000)

(Continued)

Table 2. Continued

Map ID	Bend or basin type (cf. Fig 1)	Ancient or active	Age of activity	Bend name	Country	Associated master faults	Selected references
58	Gentle restraining bend	Active	L. Quaternary	Transverse Ranges (San Geronio Pass)	USA (California)	Eastern California	Spotila & Anderson (2004)
59	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Unnamed	Venezuela	Bocono	Schubert (1980)
60	Gentle restraining bend	Active	L. Quaternary	Unnamed, near Bam 2004 epicentre	Iran	Gowk–Sabzevaran	Talebian <i>et al.</i> (2004)
61	Gentle restraining bend	Active	L. Quaternary	Various restraining bends	Iceland	Selsund	Angelier <i>et al.</i> (2004)
62	Gentle restraining bend	Active	Late Neogene–L. Quaternary	Wairau ‘big bend’	New Zealand	Wairau Fault	Yeats & Berryman (1987); Berryman <i>et al.</i> (1992)
63	Gentle restraining bend	Active	L. Quaternary	Zahret el-Qurein (Zahret Qurein)	Jordan	Jordan Valley	Garfunkel (1981)
Ancient restraining bends							
64	Sharp restraining bend	Ancient	Cenozoic–Palaeocene	Cascade Mountains	USA (Washington)	Ross Lake	Miller (1994)
65	Sharp restraining bend	Ancient	Early Cenozoic	Owl Creek	USA (Wyoming)	North Owl Creek; Shotgun Butte	McClay & Bonora (2001)
66	Sharp restraining bend	Ancient	Tertiary	Unnamed	Netherlands	Unnamed	Dooley <i>et al.</i> (1999)
67	Gentle restraining bend	Ancient	Not given in paper	Anninghe	China	Xianshuihe; Zemuhe	Swanson (2005)
68	Gentle restraining bend	Ancient	Tertiary; Palaeogene	As Pontes Basin (Ebro Basin)	Spain	Penedes Valles; Gandesa–Ulldemolins; El Camp; Pont d’Armentera	Anadon <i>et al.</i> (1985)
69	Gentle restraining bend	Ancient	Palaeogene	Black Hills	USA (Alaska)	Beringian	Worrall (1991)
70	Gentle restraining bend	Ancient	Late Precambrian	Brazilion craton	Brazil	Pabos	Corsini <i>et al.</i> (1996)
71	Gentle restraining bend	Ancient	Late Palaeozoic	Casco Bay	USA (Maine)	Norumbega; Flying Point	Swanson (2005)

72	Gentle restraining bend	Ancient	Devonian–Carboniferous	Cobequid Highlands	Canada	Rockland Brook; Cobequid	Miller <i>et al.</i> (1995)
73	Gentle restraining bend	Ancient	Eocene?	Gasherbrum Range	Pakistan	Karakoram	Searle & Phillips (2005)
74	Gentle restraining bend	Ancient	Ordovician–Devonian	Gaspé Belt	Canada	Lac Cascapedia; Mont Albert; Shickshock Sud	Sacks <i>et al.</i> (2004)
75	Gentle restraining bend	Ancient	Mesozoic–Cenozoic	Jargaland	Mongolia	Haar Us Naar	Cunningham <i>et al.</i> (1996a); Howard <i>et al.</i> (2003); Cunningham (2007)
76	Gentle restraining bend	Ancient	Cretaceous–Tertiary	Magallanes	Argentina	North Scotia Ridge	Cunningham (1995); Lodolo <i>et al.</i> (2003)
77	Gentle restraining bend	Ancient	Mesozoic–Cenozoic	Mt Tormeno	Italy	Tormeno; Gamonda	Zampieri & Massironi (2005)
78	Gentle restraining bend	Ancient	Permian–Tertiary	Quealy Dome	USA (Wyoming)	South Quealy	Stone (1995); McClay & Bonora (2001)
79	Gentle restraining bend	Ancient	Cenozoic	Unnamed	Iran	Purang	Walker & Jackson (2004)

Green, pull-apart basins or releasing bends. Red, restraining bends.

In this paper, I reserve *transform fault* only for strike-slip faults in oceanic crust that offset and abruptly transform motion into orthogonal oceanic spreading ridges as originally defined by Wilson (1965; Figs 1 & 2; Table 1). Yeats *et al.* (1997) discuss some of the problems in the wider application of the term *transform fault* to continental strike-slip faults like the San Andreas and Sumatra faults, where strike-slip motion is not abruptly transformed at their ends on to other orthogonal structures, and where the strike-slip fault may only accommodate only a fraction of the total plate motion. Because of the problems of the application of the term *transform* to continental settings, I use the general term *strike-slip fault* for sub-aerial strike-slip faults in continental settings at any scale, whose slip vectors are roughly parallel to their strike (Sylvester 1988; Fig. 1). A *strike-slip fault system* has a number of kinematically related and subparallel strike-slip faults typically within a broad plate-boundary zone up to several hundred kilometres in width (e.g. the San Andreas strike-slip system). Two commonly used terms also used in this review are *transpression* and *transtension*, broadly defined as steep, strike-slip-influenced deformation zones that deviate from simple shear by a component of shortening (transpression) or extension (transtension) across the zone (Sanderson & Marchini 1984; Fossen & Tikoff 1998).

Following Sylvester (1988) and Yeats *et al.* (1997), I avoid using the words *wrench* or *transcurrent*, which were widely used in pre-plate tectonic literature and in articles and books about petroleum exploration in strike-slip fault settings (cf. Wellman 1955; Moody & Hill 1956; Wilcox *et al.* 1973; Freund 1974; Sylvester 1984; Harding *et al.* 1985; Lowell 1985; Mount & Suppe 1987; Jamison 1991; Zolnai 1991; Milnes 1994). Both *wrench* and *transcurrent* were originally defined in the early 20th century with genetic and/or dynamic connotations (cf. Sylvester 1988, for a historical review and the modern usage problems of both terms).

Releasing-bend terms used in this paper

For strike-slip bends of an extensional character (right-stepping offset for a right-lateral fault, left-stepping offset for a left-lateral fault, Fig. 1b), I will use the terms that Crowell (1974a, b) developed for structures along the San Andreas fault zone. These terms include *releasing bend* and *releasing double bend* (Crowell 1974a) and *pull-apart basin* (Burchfiel & Stewart, 1966; Mann *et al.* 1983) that are used interchangeably in this paper (cf. fig. 1 of Cunningham & Mann 2007; Figs 1b, c & d). The two approximately parallel strike-slip faults bounding pull-aparts or push-ups are simply referred to as *master faults* (Rodgers 1980). The distance

between master faults is their *fault separation* or pull-apart basin width; the distance of overlapping master faults is their *fault overlap* and roughly defines the length of the pull-apart basin (Fig. 1b). *Pull-apart* is sometimes used in petroleum geology literature to refer to rift basins produced by continental breakup (cf. Lowell 1985), but the strike-slip usage that I will follow has precedence.

Active pull-apart basins are easily recognizable along active strike-slip faults, because the basins form topographically low, fault-bounded depressions that are commonly the sites of marine embayments, inland lakes and closed topographic depressions, which in the case of the Death Valley pull-apart (−85 m b.s.l.), the Dead Sea pull-apart (−411 m b.s.l.), and multiple pull-aparts in the southern Imperial Valley of California and northern Mexico (−40–65 m b.s.l.). Therefore pull-aparts have some of the lowest, sub-aerial topographic elevations in continental areas (Figs 3–4; Table 2). Crustal thinning produced by localized extension within the stepover region localizes rapid deposition of clastic sediments eroded from surrounding highlands (Hempton & Dunne 1983) as well as intrusive, volcanic and geothermal activity at the surface (Aydin *et al.* 1990) or intercalated within the basin sediments (Einsele 1986). Pull-apart depressions of a variety of sizes form major discontinuities along otherwise straight segments of the main ‘through-going’ strike-slip fault, or ‘principal displacement zone’ (Tchalenko & Ambraseys 1970; Figs 3 & 4; Table 2).

Other synonyms for pull-apart basins are *gaps* (Quennell 1956; 1958) *tectonic depressions* (Clayton 1966), *wrench grabens* (Belt 1968), *rhomb grabens* (Freund 1971; Aydin & Nur 1982; Bahat 1983; Heimann *et al.* 1990), *dilational fault jogs* (Sibson 1985, 1986a,b, 1987), *extensional duplexes* (Woodcock & Fischer 1986; Swanson 2005), *sidewall basins* (Gibbs 1989), *tensile bridges* (Gamond 1987), *stepover basins* (Wakabayashi *et al.* 2004, 2007), and *dilational stepovers* (Oglesby 2005). T. Dooley (pers. comm., 2006) uses the term *in-line depressions* as basins along strike-slip faults that are formed between Riedel shears that amplify as an anastomosing fault develops by R- and P-shear linkage (cf. the *tensile bridges* of Gamond 1987).

Rhombchasms were defined by Carey (1958), in his early outline of global plate tectonics, as a special case of parallel-sided pull-apart basins that are floored by oceanic crust. *Intra-transform spreading centres* were proposed by Fornari *et al.* (1989) as a special case of pull-aparts on oceanic transform faults that have developed short spreading centres and are floored in part by oceanic crust. *Sag ponds*, which are common along continental strike-slip faults may be related to the

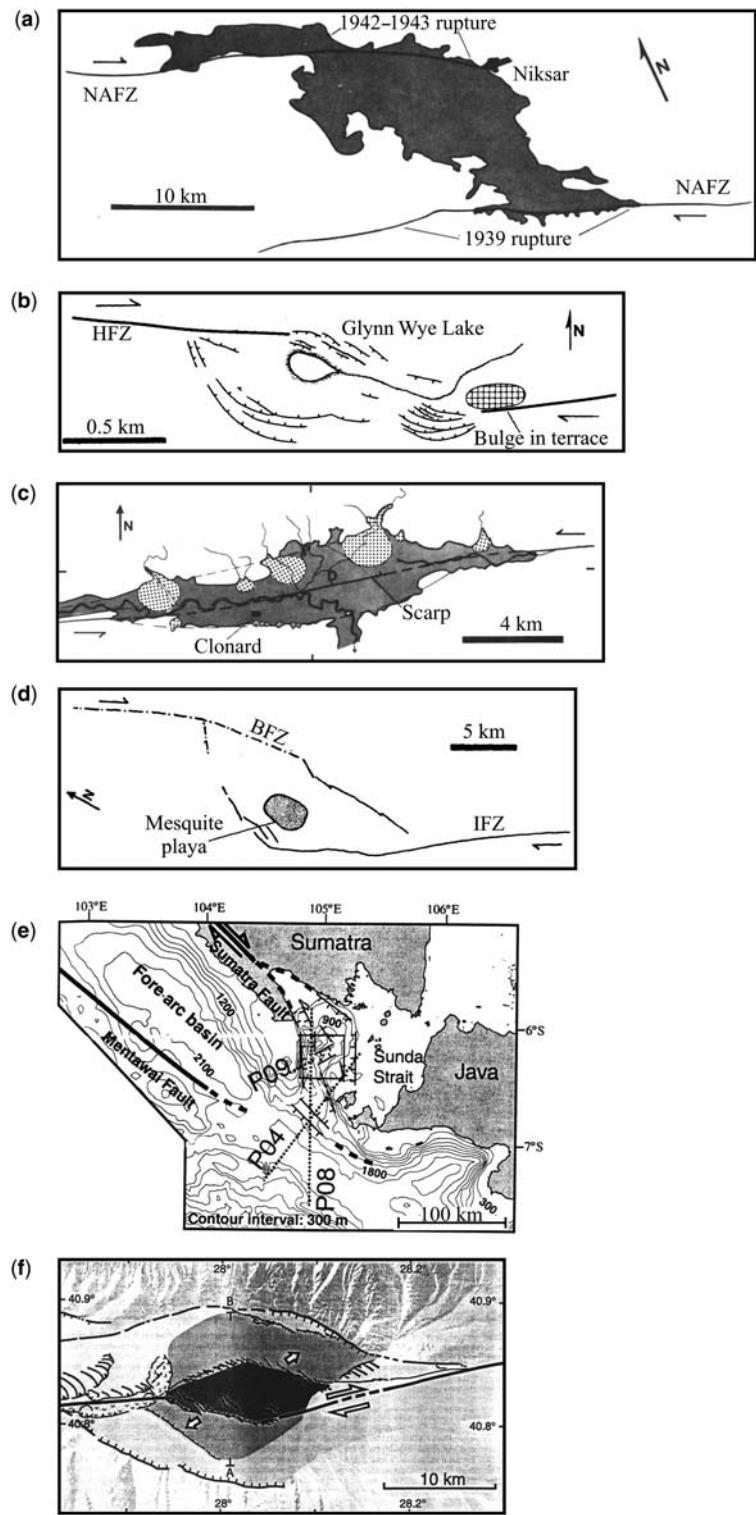


Fig. 3. See caption on p. 36.

formation of small, underlying pull-apart basins by subtle, small shifts in the trace of the fault by metres to hundreds of metres (Sylvester 1988).

Pull-apart basins are one type in a family of small but potentially thick, strike-slip-controlled basins that include *fault-wedge basins* (Crowell 1976; Biddle & Christie-Blick 1985), *fault-angle depressions* (Ballance & Reading, 1980), *ramp basins* (P. Mann *et al.* 1991) or *push-down basins* (Cobbold *et al.* 1993), and *fault-flank depressions* (Crowell 1976). The characteristics and tectonic controls on these other types of strike-slip basins are reviewed by Biddle & Christie-Blick (1985), Cobbold *et al.* (1993), Ingersoll and Busby-Spera (1995), and Allen & Allen (2001).

Fault termination basins

Fault termination basins or *single releasing bends* in the terminology of Crowell (1974a, b) are an important class of strike-slip basin that warrants illustration in Figure 5 because of their similarity to and possible confusion with classical pull-aparts as illustrated in Figures 3–4. These basins terminate strike-slip displacement along a single strike-slip fault or fault zone by diffuse and splayed extensional structures in a characteristic ‘horsetail’ pattern that ranges in width from about 30 km for the sub-aerial, Neogene La Tet Fault in the Pyrenees Mountains of Spain (Cabrera *et al.* 1988; Fig. 5a), to larger examples along the strike-slip Median Tectonic Line in Japan (Itoh *et al.* 1998); along the Motagua strike-slip fault of northern Central America (Guzmán-Speziale 2001); along the North Island strike-slip faults of New Zealand (Mouslopoulou *et al.* 2007); and to one extreme example over 1000 km in width that Briais *et al.* (1993) proposed for the offshore termination of the Red River strike-slip fault zone in the South China Sea (Fig. 5b). These extensional features are confined to one side of a single strike-slip fault, and therefore differ in character from a localized extension in a pull-apart basin confined by two, subparallel, master strike-slip faults (Fig. 1b). Fault termination features are not discussed

further in this paper and are not included as part of the bend compilation shown on Figure 2.

Duplex terminology

In this paper, I do not use the *duplex terminology* for strike-slip bends that was adapted from thrust fault terminology by Woodcock & Fischer (1986). In this terminology, *ramps* are synonyms for strike-slip bends; *straights* are synonyms for strike-slip faults subparallel to the regional slip vector; *extensional duplex* is a synonym for pull-apart basin; *contractional duplex* is a synonym for restraining bend; and *trailing contractional* or *extensional imbricate fans* are synonyms for horsetail-type fault splays at the ends of strike-slip faults.

The duplex terminology is used for outcrop studies of strike-slip fault bends (Swanson 1989a, 2005; Cruikshank *et al.* 1991; Laney & Gates 1996), but is also widely used for descriptions of larger-scale bends (Swanson 2005; Cunningham 2007; Gomez *et al.* 2007). The Crowell (1974a, b) and Woodcock & Fischer (1986) terminology for bend structures are both illustrated and compared on Figure 1 of Cunningham & Mann (2007).

Restraining-bend terms used in this paper

Frequently used terms include *gentle restraining bend* and *sharp restraining bend* (Crowell 1974a; Fig. 1c). Active restraining bends are also easily recognizable, because they localize topographically elevated ridges or folded mountain belts and expose fault-bounded, elongate belts of deeper crustal rocks (Karig 1979; Cunningham 1995; Roeske *et al.* 2003; Cunningham *et al.* 2007) that include blueschist-facies metamorphic rocks in some localities (Mann & Gordon 1996; Fig. 1c; Table 2).

Sharp restraining bends are rhomboidal in shape, are aligned along the general trend of major strike-slip faults, and range in scale up to several kilometres in length and up to hundreds of metres in width (Fig. 6b & c; Table 2). The areas of sharp restraining bends are characterized by miniature fold-and-thrust belts and marked by a

Fig. 3. Active examples of recently nucleated and relatively young spindle-shaped pull-aparts, as interpreted in the basin evolutionary model of Mann *et al.* (1983). Examples are taken from several different tectonic settings or strike-slip faults that are discussed in this paper. (a) Niksar pull-apart basin along the North Anatolian fault zone, Turkey (from Seymen 1975; Hempton & Dunne 1983). (b) Glynn Wye Lake pull-apart along the Hope Fault, New Zealand (from Freund 1971; Cowan 1990). (c) Clonard pull-apart basin along the Enriquillo–Plantain Garden fault zone, southern peninsula of Haiti (from Mann *et al.* 1983, 1995). (d) Mesquite playa pull-apart basin along the Brawley–Imperial faults, Imperial Valley, California (modified from Johnson & Hadley 1976). (e) Sunda Strait pull-apart basin at the SE termination of the Sumatra fault zone between the islands of Sumatra and Java (from Legemann *et al.* 2000). (f) Central pull-apart basin along the North Anatolian fault zone in the Sea of Marmara, from Armijo *et al.* (2002). Arrows show the direction of extension inferred from en échelon normal faults.

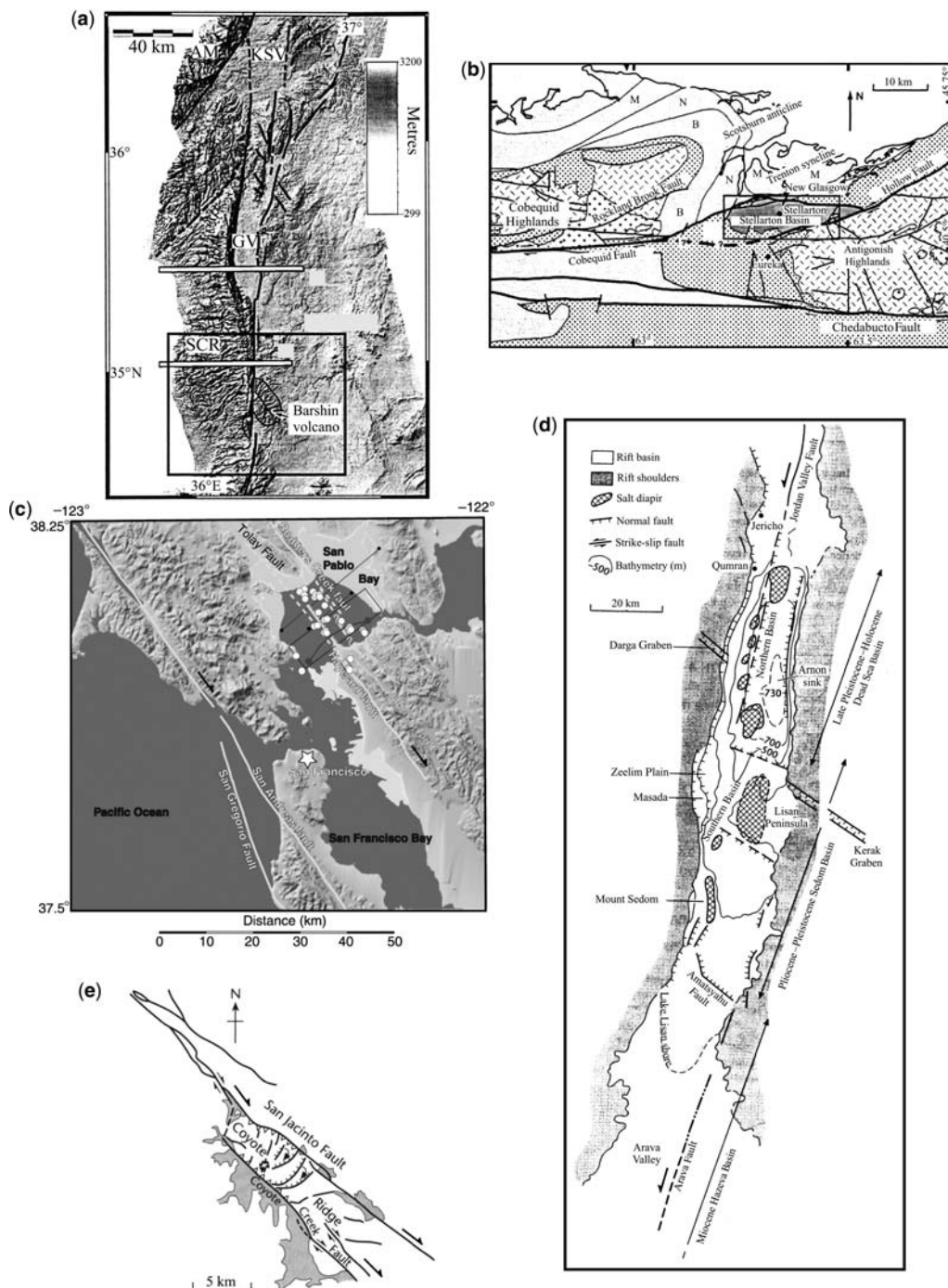


Fig. 4. Active examples of relatively older, rhomboidal pull-apart basins reflecting larger strike-slip fault offsets than those shown in Figure 3, as described in the evolutionary model of Mann *et al.* (1983). (a) Ghab pull-apart basin along the active Dead Sea fault zone (from Gomez *et al.* 2006), the Pliocene Barshin volcano is left-laterally offset by 15–20 km south of the Ghab Basin. (b) Late Palaeozoic Stellarton pull-apart basin between the Cobeguid and Hollow strike-slip fault zones, northern Appalachian Mountains, Nova Scotia (from Waldron 2004). (c) San Pablo Bay pull-apart basin between the active Hayward and Rodgers Creek fault zones (from Parsons *et al.* 2003). (d) Dead Sea Basin along the Dead Sea fault zone (from Allen & Allen 2005). (e) Coyote ridge basin along the San Jacinto Fault (from Allen & Allen (2005), based on mapping by Sharp (1975).

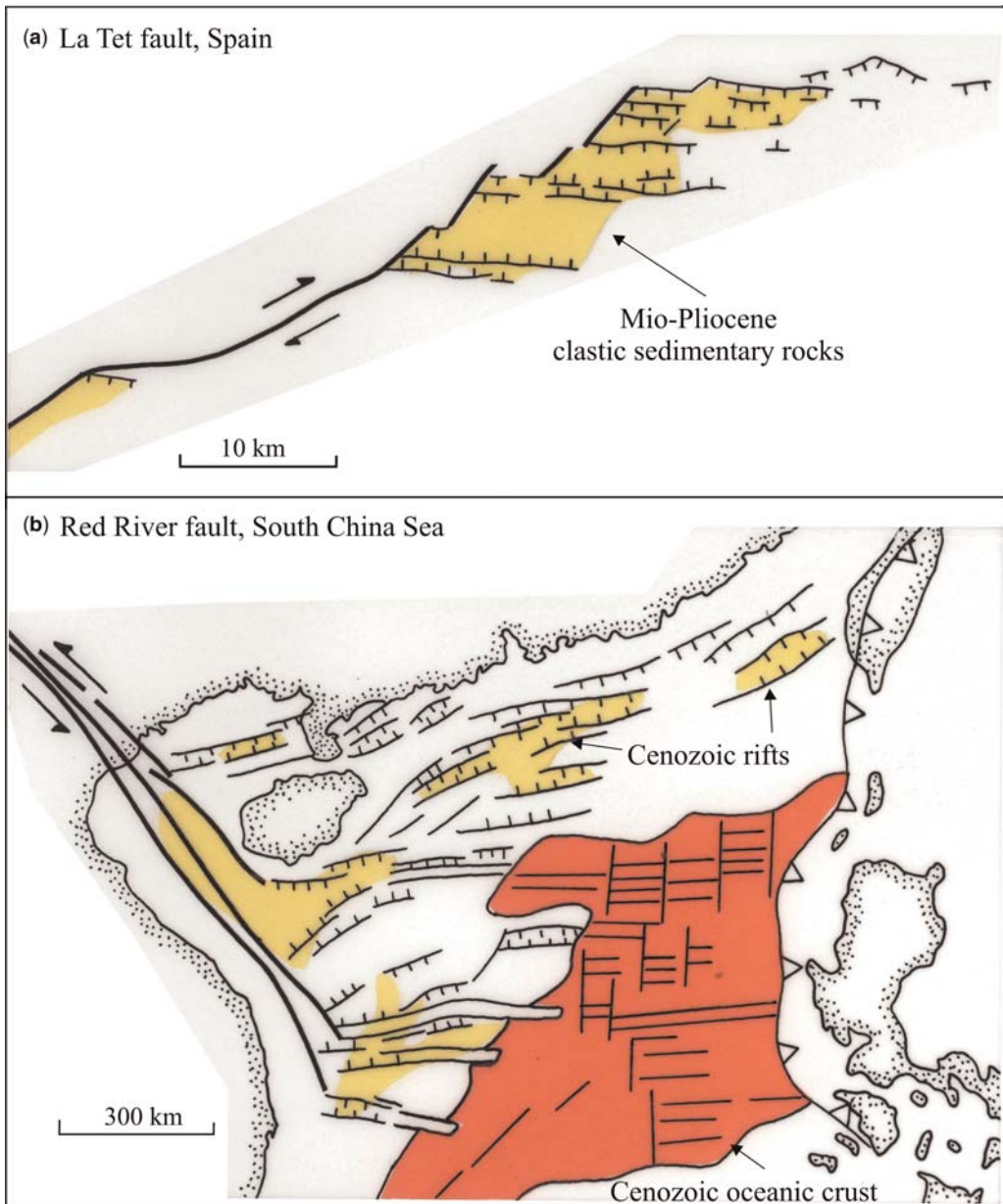


Fig. 5. Examples of strike-slip fault termination basins that are relatively common and share some characteristics with the stepover type of pull-aparts shown in Figures 3 & 4. (a) Cerdanya clastic basin (shaded yellow) formed by Late Miocene diffuse normal faulting at the termination of the La Tet strike-slip fault, Spain (modified from Cabrera *et al.* 1988). (b) Basins (shaded yellow) and area of oceanic spreading (shaded red) proposed as fault termination effects by Briais *et al.* (1993) for the Red River Fault of the South China Sea.

localized topographic uplift centred on the fold-thrust belt (Brown *et al.* 1991; Fig. 6c). *Gentle restraining bends* are more common, exhibit the characteristic lazy S and Z shapes (Fig. 6d & e)

and range in size from small features a few kilometres in width to the world's largest bends up to 200 km wide and several kilometres in elevation (Mann & Gordon 1996; Fig. 6f & g). *Paired*

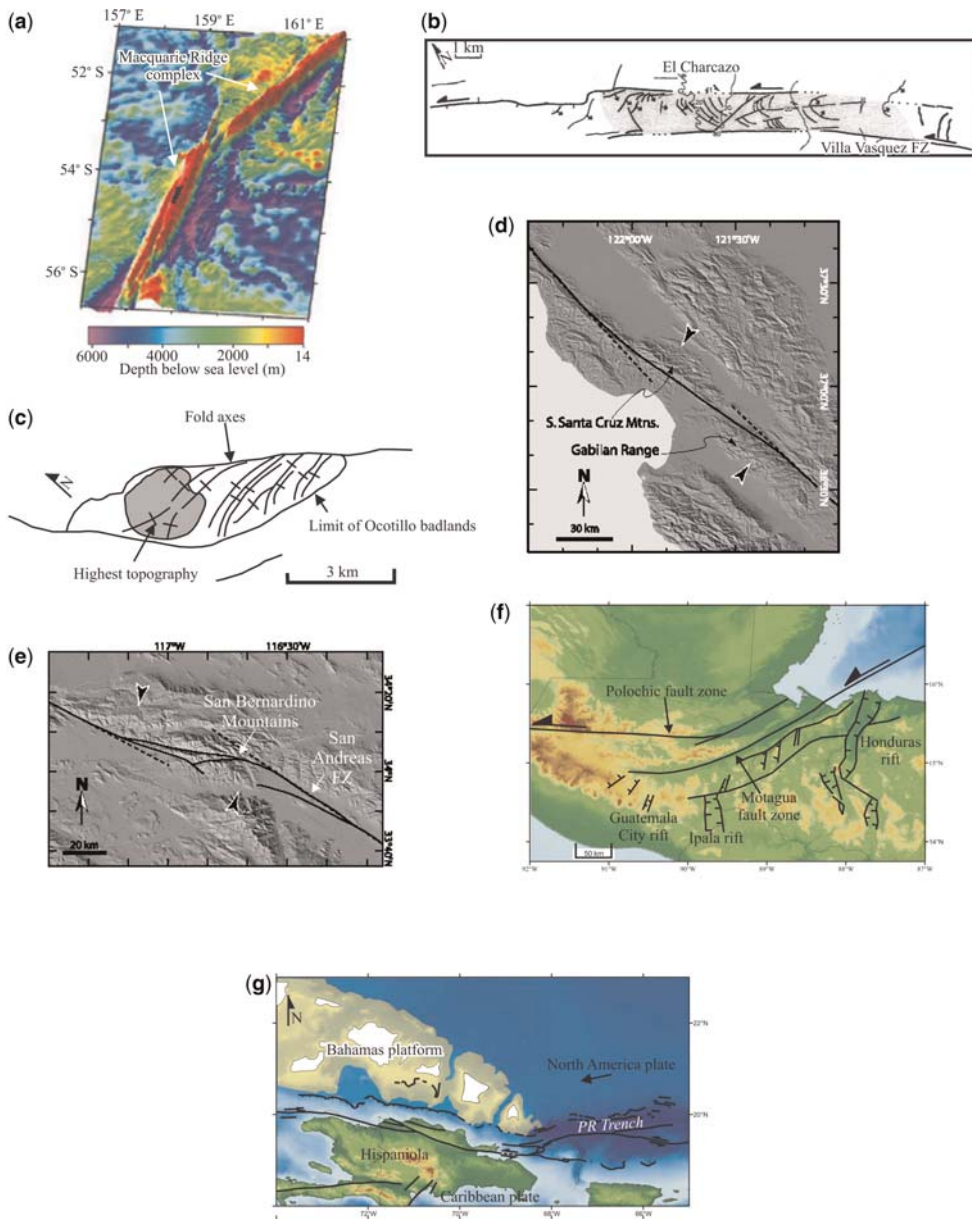


Fig. 6. (a) Macquarie Ridge from Dazcko *et al.* (2003). (b) Villa Vasquez fault zone and basin, from Mann *et al.* (1999). (c) Sketch of Ocotillo badlands, based on Segall & Pollard (1980). (d) Santa Cruz Mountains restraining bend, from Cowgill *et al.* (2004a). (e) San Bernardino Mountains restraining bend, from Cowgill *et al.* (2004a). (f) Polochic fault zone and Motagua fault zone, from Mann & Gordon (1996). (g) Bahama Bank and Hispaniola and corresponding restraining bends, from Mann *et al.* (2003).

bends are restraining and releasing bends that form adjacent to one another (Fig. 1d).

Synonyms for sharp and gentle restraining bends include *rhomb horsts* (Freund 1971; Aydin & Nur 1982; Bahat 1983); *pressure ridges* (Garfunkel

et al. 1981); *compressional stepovers* (Segall & Pollard 1980; Aydin & Nur 1985; Westaway 1995; Oglesby 2005); *push-up blocks* (Mann *et al.* 1983); *anti-dilational fault jogs* (Sibson 1985, 1986a, b, 1987; Brown *et al.* 1991); *push-up*

swells (Heimann & Ron 1987; Reches 1987); *structural knots* (Muehlberger & Gordon 1987); *compressional duplexes* (Woodcock & Fischer 1986; Gomez *et al.* 2007; Cunningham *et al.* 2007), *compressible bridges* (Gamond 1987); *pop-ups* (Stone 1995), and *restraining stepovers* (Westaway 1995; McClay & Bonora 2001; cf. figure 1 of Cunningham & Mann 2007). T. Dooley (pers. comm., 2006) uses the term *in-line uplifts* to refer to uplifts along strike-slip faults that are formed between Riedel shears that amplify as an anastomosing fault develops by R- and P-shear linkage (cf. *compressible bridges* of Gamond 1987).

Global classification of active strike-slip faults

Discussion of the settings of active releasing and restraining bends on strike-slip faults requires a global classification of active strike-slip faults. Woodcock (1986) made the first comprehensive, global classification of active strike-slip faults based on their tectonic origin and setting, which was significantly modified in the subsequent classifications by Sylvester (1988), Yeats *et al.* (1997), and Ingersoll & Busby (1995). In this paper, I will adopt Woodcock's (1986) original scheme that is summarized in Figure 2 & Table 1, but with some modifications that take into account more recent work. Sylvester (1988) adds two categories of strike-slip faults, *tear faults* and *transfer faults*, which are not discussed here since they are generally short (<50 km) and devoid of significant fault bends (cf. Escalona & Mann 2006, for a review of tear and transfer terminology and global examples).

Using Woodcock's (1986) classification, four major categories of strike-slip faults are shown schematically in Figure 2:

1. *ridge transform faults in oceanic crust* that offset spreading ridges (e.g. Siqueiros, Blanco, Clipperton transforms);
2. *boundary transform faults* that separate different plates and strike parallel to the direction of plate motion (e.g. the San Andreas Fault of California, the Alpine Fault of New Zealand, the Dead Sea Fault of the Middle East); this class includes propagating, slab-bounding transforms bounding plates in subduction to strike-slip transition areas like the Scotia and Caribbean (Bilich *et al.* 2001); this type of strike-slip fault has recently been compiled and named *STEP faults*, or *subduction–transform edge propagators* (Govers & Wortel 2005);
3. *trench-linked strike-slip faults* that accommodate the horizontal component of oblique subduction (e.g. the Atacama fault zone of Chile and the Median Tectonic Line of Japan); and
4. *indent-linked strike-slip faults* related to continent–continent collision (Red River fault zone, Altyn Tagh fault zone, North Anatolian fault zone).

In addition to these four categories from Woodcock (1986) I have added a fifth category that I will call *cratonic strike-slip fault systems* that are removed from active plate boundaries, form sharp restraining bends based on older crustal faults, and act as 'concentrators' of intraplate stresses (Talwani 1999; Gangopadhyay & Talwani 2003). In this paper, I will organize my descriptions of strike-slip fault systems into these four categories with emphasis on faults of each category that exhibit either releasing, restraining or paired bends.

Observations on the evolution of releasing bends on strike-slip faults from natural examples and models

Most previous workers have focused their investigations of pull-apart basins on some observable or measurable aspect of single natural basins, at scales ranging from 'outcrop basins' whose 'basin fill' is typically calcite or quartz, to large-scale sedimentary basins up to several hundred square kilometres in size and containing complex sedimentary facies totalling over 10 km in thickness (Hempton & Dunne 1983; Ingersoll & Busby 1995; Figs 3–4). The detailed geological mapping and description of these natural pull-apart examples at outcrop to basinal scales have provided the observational basis and parameters for continuum models such as the geological model proposed by Mann *et al.* (1983) for natural crustal-scale pull-apart basins (Fig. 7), as well as many modelling studies using laboratory analogue models (e.g. Dooley & McClay 1997) or theoretical models (e.g. Golke *et al.* 1994). In this section, I will review the main results of laboratory and theoretical studies and summarize how each type of study has furthered our current understanding of the continuum evolution of natural pull-aparts (Fig. 7).

For reasons of space I will place particular emphasis on the representative discoveries and insights made during the last two decades. Because of the numerous papers on pull-aparts, dating from the mid-1980s, this compilation (Fig. 1 & Table 2) has probably omitted many relevant papers and therefore should not be considered as a complete listing. Moreover, the compilation in Table 2 only contains crustal-scale bends, and does not include outcrop-scale pull-apart features as discussed by previous workers like Aydin & Nur

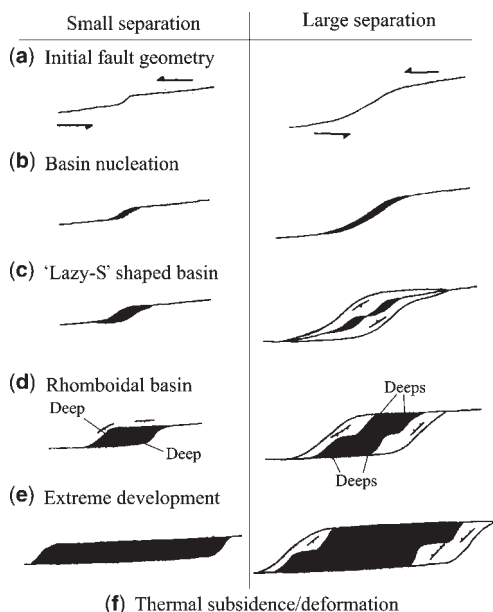


Fig. 7. Classification of pull-apart basins, as interpreted in the evolutionary model proposed by Mann *et al.* (1983). In both evolutionary sequences, a starting point of a single continuous fault with a bend is assumed, but to the left the fault bend exhibits a relatively smaller separation, or distance between the two straight fault segments. For the smaller separation case, the evolution proceeds from a nucleating basin, to a 'lazy-Z' shaped basin, to a rhomboidal basin, to an extremely elongate basin possibly floored by oceanic crust formed at a short spreading centre. For the larger separation case, two individual basins nucleate, follow the same evolutionary stages, and eventually merge to form a wider composite basin.

(1982), Bahat (1983), and Gamond (1987). In this section, I will not take into account the tectonic setting of the basins described, partly to make the fundamental point that the morphology and evolution of pull-aparts is remarkably similar, despite their four major tectonic settings proposed by Woodcock (1986) that are summarized in Figure 2 & Table 1.

Continuum of evolution in natural pull-apart basins

What causes the extensional bends on strike-slip faults?

Previously proposed models for the occurrence of solitary or arrays of pull-aparts along strike-slip faults include:

1. a nearby or changing pole of rotation produces fault curvature in the strike-slip fault, or a greater degree of transtension along the trace of the fault (Menard & Atwater 1968; Garfunkel 1981; Embley & Wilson 1992; Fig. 8a);
2. a strike-slip fault remains oblique to steady divergent motion (Fig. 8b);
3. interaction of the strike-slip fault with a pre-existing crustal heterogeneity results in a fault bend at or near their point of intersection (Mann *et al.* 1985; Mann *et al.* 2007a; Fig. 8c);
4. incompatible slip at a junction with another strike-slip fault or fault-bounded block bends the fault (Freund 1974; Bohannon & Howell 1982; Fig. 8d);
5. variable rock properties along the length of the fault will cause the fault to bend during its progressive offset and assume a sinusoidal fault trace in map view (Bridwell 1975); alternatively, fault-zone 'adherence' or frictional locking will lock the principal fault and form a new subparallel but curving splay fault (the 'sidewall ripout' of Swanson 1989a, 2005); and
6. intersection of a strike-slip fault with a zone of greater extensional or convergent strain – particularly in zones of continental collision – bends the fault at or near the point of intersection (Sengör *et al.* 1985; Fig. 8d). These mechanisms could also be responsible for forming restraining bends.

Controls of relative plate motion on pull-apart opening

For transtensional opening along the length of a fault related to a changing pole of rotation, Garfunkel (1981) proposes three possible scenarios for large-scale plate-tectonic control on the formation of pull-apart basins along a continental-boundary strike-slip fault like the San Andreas or Dead Sea strike-slip fault systems (Fig. 9).

In the simplest scenario, pure strike-slip motion occurs along some segments of the boundary that are segmented by either releasing- or restraining bends when the fault strike deviates slightly from the local strike direction (Fig. 9b). A paired bend (Fig. 1d) forms when releasing- and restraining segments form adjacent to one another and are similar in areal extent and fault separation (Fig. 9a). Fault segments of pure strike-slip motion are parallel to small circles of plate rotation about the interplate pole of rotation (thin arrows on Figure 9a point in the direction of the pole).

In a second scenario, shown in Figure 9b, the presence of many releasing- and restraining-bend segments causes the average trend of the overall

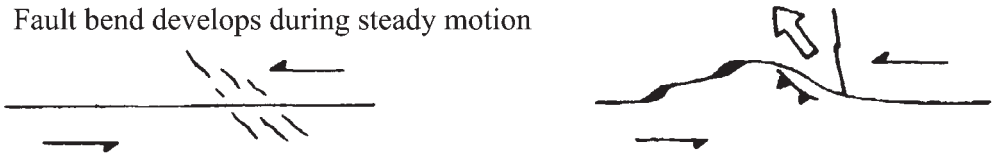
(a) Change in motion across the fault



(b) Fault oblique to steady motion



(c) Fault bend develops during steady motion



(d) Constriction of wedge during convergence

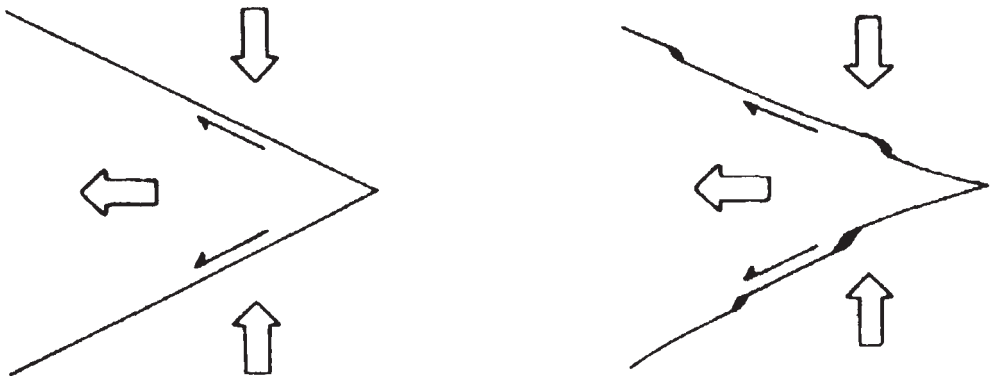


Fig. 8. Possible tectonic scenarios to explain the initiation of bends along strike-slip faults and the formation of pull-apart basins (modified from Mann *et al.* 1983). (a) Change in motion across fault. (b) Fault oblique to steady motion. (c) Fault bend develops during steady motion to create a double or paired bend consisting of adjacent restraining- and releasing-bend segments. (d) Constriction of a triangular-shaped wedge or 'escape block' during continental convergence. See text for discussion.

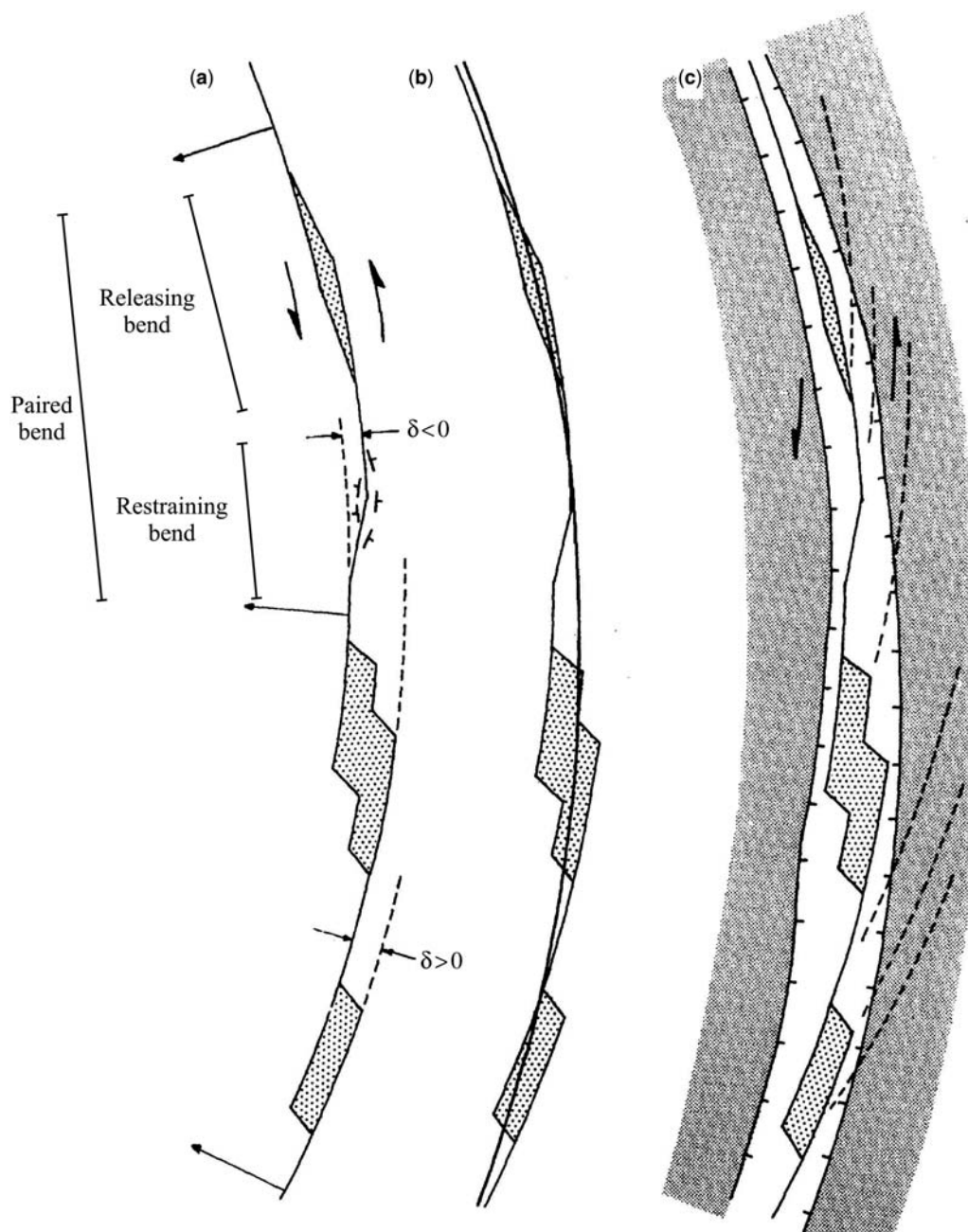


Fig. 9. Creation of pull-apart basins in a continental strike-slip fault tectonic setting, modified from Garfunkel (1981). (a) In the simplest tectonic interpretation, thin arrows point to the direction of the pole of relative motion that controls the kinematics of the strike-slip fault system. (b) Same tectonic setting as (a), with the plate boundary approximated by a circular arc that does not define either the pole of rotation or control the local fault kinematics. (c) Same situation, but with a small component of transverse motion added; the overall relative plate motions are given by the broken lines. See text for discussion.

plate-boundary fault to not accurately define the pole of rotation. In the third and most complex scenario, a component of additional transtension across the overall plate boundary results in the production of new surface area by the stretching of the plate edges and creation of a prominent fault valley that would correspond with the plate-boundary-parallel Imperial Valley along the San Andreas fault zone (Yeats *et al.* 1997) or the Dead Sea Valley along the Dead Sea fault zone (ten Brink *et al.* 1999; Fig. 9c). Overall, fault slip is resolved or 'partitioned' into a component of pure strike-slip aligned along the trend of the fault valley, and a component of orthogonal extension aligned at right angles to the trend of the fault valley.

This more complex scenario results in a complex structure and morphology: pull-apart basins embedded within a wider strike-slip valley (Fig. 9c). Pure strike-slip faults embedded in the fault valley are not necessarily parallel to the small circles of rotation (indicated by broken lines in Fig. 9c). In any or all of these scenarios, there are many potential geological complexities if the pole of rotation and plate kinematics were to suddenly change over large areas of the fault zone and the pre-existing fault traces and basins would have to adapt to the change in kinematics (ten Brink *et al.* 1999).

Parkinson & Dooley (1996) have presented a model similar to that of Garfunkel (1981) and Mann *et al.* (1983) for the predicted locations of en échelon arrays of pull-apart basins at places where the small circles of rotation obliquely intersect the trend of the plate boundary. They postulate that straight strike-slip fault segments, which are more efficient at accommodating slip than curved or irregular traces (Aydin & Nur 1985), will bidirectionally propagate until the strike of the segment is no longer compatible with the changing plate trajectory. To achieve maximum efficiency, most of the bends formed as accommodation structures are releasing bends, although some restraining bends will form as a result of non-parallel fault segments (Fig. 9a).

Nucleation of pull-apart basins

Early maps of pull-aparts by Quennell (1956, 1958), Carey (1958), and Clayton (1966) in New Zealand and along the Dead Sea fault zone were schematically shown as 'rhomb grabens' or 'sharp pull-aparts' (Crowell 1974a), with orthogonal angles between the master strike-slip faults and the normal faults accommodating basin extension. Careful field mapping by Freund (1971), and more recently by Cowan (1990), of active fault traces of the right-lateral Hope fault zone of New Zealand, shows that the basin-bounding master faults can

be non-parallel in strike and appear as 'spindle' (Mann *et al.* 1983) or 'almond' (Dooley & McClay 1997) shapes (Fig. 3b–d & f) on the map rather than the classical, rectilinear rhomboidal shapes (Fig. 4a–e).

If the two master faults 'underlap', then they are connected by a short, oblique fault segment that makes a 10–15° angle with the master faults as seen in pull-aparts at Glynn Wye Lake on the Hope Fault (Clayton 1966; Freund 1971; Cowan 2000; Fig. 3b), the Clonard Basin on the Enriquillo–Plantain Garden fault zone in the northern Caribbean (Mann *et al.* 1983, 1995; Fig. 3c), Mesquite Playa on the Brawley and Imperial fault zones in southern California (Weaver & Hill 1978; Fuis *et al.* 1982; Fig. 3d), and the Central Basin on the North Anatolian fault zone in the Sea of Marmara (Fig. 3f) (Armijo *et al.* 2002).

This initial master-fault geometry observed for young basins is similar to the *gentle restraining-bend* geometry proposed by Crowell (1974a; Fig. 1b). The nucleating basins may be marked by obliquely striking faults at a discontinuity along the master faults that are not associated with subsiding basins (e.g. en échelon faults in the Tiburon area of the Enriquillo–Plantain Garden fault zone in SW Haiti–Mann *et al.* 1983, 1995; en échelon faults at the Laoqianning–Kangding fault junction on the Xianshuihe fault zone, Allen *et al.* 1991). When the width of the basin (or spacing between master faults) is relatively small (<5 km), the young basins have a characteristic 'spindle shape' morphology, as seen, for example, at Glynn Wye Lake, New Zealand; Clonard Basin, Haiti; Mesquite Playa in southern California; and the Central Basin in the Sea of Marmara, Turkey (Fig. 3a). Non-parallelism of master faults leads to space problems at the edge of the pull-apart basin, problems that are expressed morphologically by bulging, folding and uplift of Quaternary terraces as observed at Glynn Wye Lake on the Hope Fault (Clayton 1966; Freund 1971; Cowan 1990; Fig. 3b).

Mann *et al.* (1983) have interpreted these spindle-shaped pull-aparts to represent the initial stage in a continuum model for pull-apart development, as shown schematically in Figure 1b & Figure 7a. Because of the limited amount of opening, the structural depression overlying nucleating pull-aparts illustrated on Figure 7a–f is generally not as prominent as the more evolved rhomboidal pull-apart basin types illustrated on Figure 3b–d. Many nucleating pull-aparts, including those shown in Figure 3a–d & f are above sea-level, sub-aerial, or covered by small, shallow lakes, and exhibit low-relief fault scarps that localize only gradual facies changes across the depression (Hempton 1983; Hempton & Dunne 1983; Link *et al.* 1985).

Lazy S- and Z-shaped pull-apart basins

Following nucleation of pull-aparts at releasing bends with little or no fault overlap (Fig. 7a), continued strike-slip offset produces basin shapes that Mann *et al.* (1983) colloquially called *lazy-S shape* for pull-aparts between left-lateral faults (and *lazy-Z shape* for those between right-lateral faults (cf. the shapes of the Niksar and Sunda basins in Fig. 3a & e). These basin shapes may be modified by local deformation associated with lengthening non-parallel master faults, as seen in the Bichuan Basin of China (Deng *et al.* 1986).

Mann *et al.* (1983) proposed that the lazy S- and Z-shapes of pull-apart basins represent a transitional stage between incipient spindle-shaped basins between master faults with no overlap and rhomboidal basins or 'rhomb grabens' between overlapping master faults (Fig. 7a). Active examples of lazy S- and Z-shaped pull-aparts have been identified in many diverse tectonic and sedimentary settings, including the mountainous region of the northern Andes of Colombia and Venezuela (Schubert 1980; Velandia *et al.* 2005); the deeply sedimented Guinea strike-slip margin of the Equatorial Atlantic (Benkhelil *et al.* 1995); and the sparsely sedimented Blanco oceanic transform fault in the Pacific Ocean (Embley *et al.* 1987).

Larger fault separations, as shown in Figure 7, lead to more prominent S- and Z-shaped basins, along with the possibility of dual, coalescing basins formed in the stepover area, as proposed by Aydin & Nur (1982) and observed from subsurface studies in the Salton Trough of southern California (Fuis *et al.* 1982); the Mor Basin within the Dead Sea Basin (Bartov & Sagy 2004); the Bohai composite pull-apart basin of eastern China (Allen *et al.* 1997; Allen 1998); and the Bichuan Basin in western China (Deng *et al.* 1986).

Increased extension across S- and Z-shaped basins along normal to oblique-slip normal faults at 35–50° to the master faults results in a prominent topographic and sedimentary basin bounded by prominent normal faults at the flanks of the basin. This stage of pull-apart is typically a deep depression surrounded by uplifted and deeply eroded highlands of basement rocks. For example, Death Valley, a prominent Z-shaped pull-apart basin of Miocene–Recent age, is bounded by low-angle normal faults (Burchfiel & Stewart 1966; Biddle & Christie-Blick 1985; Serpa 1988; Cowan *et al.* 2003). The floor of the basin is 85 metres below sea-level, and about 3 km lower than the peaks in the surrounding mountains. The neighbouring Panamint and Saline valleys are geomorphically similar pull-aparts also bounded by low-angle normal faults (Burchfiel *et al.* 1987). The world's largest S-shaped pull-apart basin is Lake Baikal in

Russia, covering over 31 494 km² and exhibiting 4477 m of structural relief between the deepest part of the lake (–1637 m) and the adjacent peaks (2840 m) (Petit *et al.* 1996; Lesne *et al.* 1998). Because of the large amount of relief, sedimentation in large S- and Z-shaped basins like Death Valley and Lake Baikal is highly asymmetrical and consists of alluvial-fan deposits localized along fault scarps forming the sides of the basin (Hempton 1983; Hempton & Dunne 1983). Proximal fan deposits grade into distal sand and mud facies on the flat basin floor.

Rhomboidal pull-apart basins

Pull-aparts with increased opening and offset along the strike-slip master faults evolve from lazy Z and S shapes to assume the more familiar rhomboidal or 'rhomb graben' appearance (Fig. 7d) and exemplified in the representative rhomb-shaped pull-aparts compiled on Figure 4a–e. In the terminology of Crowell (1974a, b), rhomboidal pull-aparts would correspond with his *sharp pull-aparts* (Fig. 1b). In most cases, basin lengthening to rhomboidal shapes does not involve basin widening; however, coalescence of adjacent basins may form larger composite pull-aparts with greater fault separations, as proposed by Aydin & Nur (1982) and shown schematically in Figure 7d & e.

Rhomboidal basins are longer (greater master fault overlap), wider (greater fault separation), and form deeper topographic basins than less-evolved spindle-shaped and S- and Z-shaped basins (Fig. 7). A good example of a rhomboidal basin is the Dead Sea pull-apart basin, whose 8–10-km-wide basin floor is –411 m b.s.l. and is filled by more than 10 km of clastic and evaporitic sedimentary rocks (Garfunkel & Ben-Avraham 1996; Fig. 4d).

As in spindle-shaped basins, non-parallelism of master faults leads to bulging of terraces, topographic uplift, and localized inversion of basinal sediments along the misaligned basin edge in the rhomboidal Hanmer pull-apart of New Zealand (Wood *et al.* 1994). Hempton & Dunne (1983) compiled stratigraphic thickness information from 20 modern and ancient basins of spindle, S and Z, and rhomboidal shapes, to conclude that the stratigraphic thickness and asymmetry increases with fault overlap (i.e. rhomboidal basins contain thicker basinal fills than spindle or S- and Z-shaped basins). In general, the master faults and transverse normal faults that create the subsidence in pull-aparts are not exposed at the surface, and must be imaged using the seismic-reflection techniques that are discussed below. Notable exceptions to this rule are the Coyote Ridge rhomboidal pull-apart basin along the San Jacinto fault zone in

southern California (Sharp 1975) and an unnamed rhomboidal pull-apart between the Xianshuihe and Ganzi–Yushu fault zones of SW China (Allen *et al.* 1991). For both examples, orthogonal normal faults within the overlap area of the master faults are well exposed.

An interesting geological characteristic observed in several rhomboidal pull-apart basins is the presence of basinal deeps at the distal ends or ‘inside corners’ of the basin separated by a shallow sill, as shown schematically in Figure 7d. The deeps are roughly circular and may or may not be fault-bounded; they are generally distributed diagonally across the floor of the basin. Examples of rhomboidal pull-aparts with twin, diagonal deeps include the Mirogoane Basin of Haiti (Mann *et al.* 1983); the Vienna Basin of Austria (Royden *et al.* 1982; Hinsch *et al.* 2005); the Elat, Aragonese and Dakar basins of the Gulf of Aqaba (Elat) (Ben-Avraham *et al.* 1979); the Cariaco Basin of Venezuela (Schubert 1982; Jaimes 2003); the Ghab Basin of Syria (Brew *et al.* 2001) and the Jianchuan–Hequin Basin on the Xiaojinhe fault zone in western China (Deng *et al.* 1986).

Inversion of rhomboidal basins

Examples of partially inverted rhomboidal pull-apart basins commonly deformed as a result of non-parallel master faults include the Hanmer pull-apart along the Hope fault zone of New Zealand (Wood *et al.* 1994); the offshore Dagg Basin along the Alpine fault zone of New Zealand (Barnes *et al.* 2001, 2005); and the Late Palaeozoic Stellarton Basin of Nova Scotia (Waldron 2004). Reversal of slip along master faults and subsequent rhomboidal pull-apart deformation has been proposed in the Cenozoic Catacamas pull-apart basin of Honduras (Gordon & Muehlberger 1994) and the Cenozoic Bovey Basin of SW England (Holloway & Chadwick 1986).

Migration of depocentres within rhomboidal pull-apart basins

Outcrop-based mapping by Zak & Freund (1981), Wakabayashi *et al.* (2004), and Wakabayashi *et al.* (2007) and subsurface mapping using wells, seismic-reflection and gravity methods by ten Brink *et al.* (1999) and Lazar *et al.* (2006) all suggest that pull-apart depocentres do not remain stationary for periods of time from hundreds of thousands to millions of years, but instead migrate along the strike of the master faults that control them. In the Dead Sea pull-apart, Zak & Freund (1981) first noted that the depression is occupied by three distinct Miocene to Recent sedimentary basins whose depocentres are displaced northward

over a lateral distance of 50 km with time. They postulated that the thinned, basinal area of the Dead Sea was fixed to the Levant plate and was detached from and lagging behind the relative northward motion of the thicker Arabian plate. Ten Brink & Ben-Avraham (1989) postulated that migration of the depocentre over a master-fault-parallel distance of 50 km reflects the directional propagation of the tips of both master faults; this propagation results in the northward migration of depocentres recognized by Zak *et al.* (1981). Lazar *et al.* (2006) have revised these previous models in their sequential basins model that includes a number of active sub-basins formed between en échelon strands of the main plate boundary fault. These basins become larger and deeper as motion progresses, and are bound by transverse normal faults.

Along the strands of the northern San Andreas fault zone, Wakabayashi *et al.* (2004; 2007) relates a ‘wake’ of deformed and inverted pull-apart or fault valley sediments to localized inversion effects similar to those described for slightly non-parallel master faults of the Hope fault zone of New Zealand (Freund 1971; Wood *et al.* 1994; Cowan *et al.* 2000). The net effect of this migrating depocentre model is for the active pull-apart to show less structural relief and stratigraphic thickness than predicted for the amount of total offset on the master faults.

Extreme pull-apart basin development

With continued strike-slip offset and minimal inversion effects, rhomboidal pull-apart basins may lengthen indefinitely into long, narrow oceanic basins, whose basin length, or master-fault overlap, can exceed basin width, or fault separation, by a factor of 10. The Cayman Trough, a pull-apart basin dominating the North America–Caribbean strike-slip system, is a unique example of an active pull-apart basin which has evolved until its master-fault overlap of approximately 1400 km greatly exceeds its fault separation of 100–150 km. Studies of magnetic anomalies in the floor of the Cayman Trough suggest that its spreading ridge initiated between the Early Eocene (Leroy *et al.* 1996; 2000) and the Mid-Eocene (Rosencrantz *et al.* 1988).

Other extreme pull-aparts in continental-margin settings occupied by short spreading ridges and underlain in part by oceanic crust include one of the five pull-aparts along the Blanco transform in the Pacific (Embley & Wilson 1992); an array of five pull-aparts in the Andaman Sea (Curray 2005); and four pull-aparts in the Gulf of California (Einsele 1986). Lallemand & Jolivet (1986) proposed that the oceanic crust of the Sea of Japan originated as an extreme, oceanic-floored pull-apart

formed by oblique motion in a back-arc setting in Miocene to Recent times.

Models for pull-aparts based on heat flow and thermal modelling

Because of the small size of most pull-aparts, localized heat formed during crustal extension is quickly dissipated laterally through the colder basin walls by conduction (Bradley 1983) and by convection through fluid migration along dilatant fault planes (Sibson 1985; 1986a, b; 1987; Fig. 7f). Narrower basins formed by limited fault separation between master faults would also cool quickly. Because of their rapid cooling, pull-aparts would exhibit more rapid subsidence than predicted by the uniform extension model applicable to stretched continental margins (Pitman & Andrews 1985; Allen *et al.* 2001).

More rapid subsidence of pull-aparts related to lateral heat loss could contribute to deep, sediment-starved basins like the Cayman Trough of the Caribbean (Rosencrantz *et al.* 1988) and the Ridge pull-apart basin of southern California (Crowell & Link 1982; May *et al.* 1993) which are deepening faster than the rate of sedimentary infilling (Rosencrantz *et al.* 1988). Bradley (1983) and Mann *et al.* (1983) interpreted a two-stage sedimentary history for the 250 000 km² Magdalena Basin that conforms well to predictions based on crustal stretching models. Stretching of the lithosphere within the rhomboidal pull-apart was accompanied by crustal thinning and increased heat flow. Following the rift phase, recovery of isotherms caused slower subsidence of the enlarged basin that buried the pull-apart.

Complexly faulted pull-apart basins

The Z-shaped Niksar (Tatar *et al.* 1995); the rhomboidal Erzincan (Barka & Gulen 1989), pull-apart basins along the North Anatolian fault zone, and the rhomboidal Hula Basin on the Dead Sea transform (Heimann & Ron 1987; Zilberman *et al.* 2000), show that these basins are intersected by other active faults, some of which are not parallel to master faults assumed by previous workers such as Aydin & Nur (1982), Mann *et al.* (1983) and Hempton & Dunne (1983). Therefore, their morphology, continued opening and deformation are not simply the result of the simple pull-apart opening model shown in Figure 7 but are likely to reflect more complex tectonic origins.

Extinction of rhomboidal pull-apart basins

Zhang *et al.* (1989) described an array of four active rhomboidal pull-apart basins along a 100-km-long

segment of the Haiyuan fault zone in NW China. Several of these basins exhibit the development of a new strike-slip fault along one of the basin-margin master faults (with a normal component of throw) or diagonally crossing the centre of the Quaternary basin fill. These new diagonal faults on the basin floor differ from those oblique basinal faults described from spindle-shaped basins as shown in Figure 3b–d & f, because in the case of the Haiyuan basins they occur in rhomboidal basins with 5–10 km of master fault overlap. Zhang *et al.* (1989) concluded that the migration of the master strike-slip faults toward the center of the basin would lead to the end of extension across the basin; a decrease in the areal size of the pull-apart basin; and thereby would result in the process of 'pull-apart extinction'. Zhang *et al.* (1989) speculated that the likely mechanism for this process was the tendency of the arcuate Haiyuan fault zone to straighten itself.

Examples of rhomboidal pull-aparts with documented, diagonal strike-slip faults that ruptured seismogenically across the Quaternary floor of the basin in the manner described by Zhang *et al.* (1989) for the Haiyuan fault zone, include a 1951 event that ruptured the rhomboidal Beng Co pull-apart basin in Tibet (Armijo *et al.* 1986); a 1966 event that ruptured the rhomboidal Cholame pull-apart along the San Andreas fault zone (Brown *et al.* 1967); and the 1992 Landers event that ruptured the Homestead Valley basin between two parallel master faults (Sowers *et al.* 1994).

Hurwitz *et al.* (2002) interpreted a grid of seismic-reflection data from the rhomboidal Sea of Galilee pull-apart, and concluded that motion on one master fault on the western side of the basin has ceased, with all motion now concentrated on a single master fault. This implies that stretching of the Sea of Galilee basin floor has ceased and that the pull-apart phase of the basin history has ended.

Preservation potential of active and ancient pull-apart basins

Aydin & Nur (1982) tabulated information on 64 active and ancient pull-aparts and concluded that most were rhomboidal, with length to width ratios of about 3:1. Subsequent work has shown that some of the measured populations were made from other types of strike-slip basins and not pull-aparts as defined above (i.e. compare their table 1 with the pull-aparts listed in my Table 2). Nevertheless, their observation supports the larger number of rhomboidal pull-aparts (Fig. 4a–e) than either spindle-shaped basins (Fig. 3 b–d & f), S- or Z-shaped basins (Fig. 3a & e), or extreme rhomboidal basins (the Cayman Trough), as compiled in my Table 2.

One explanation for the larger number of rhomboidal basins is the tendency for older, longer basins with higher ratios to either become extinct by the inversion processes described above or to be destroyed by a sudden change in the strike-slip or overall tectonic environment (Mann *et al.* 1983). In a global study of the longevity of sedimentary basin types, Woodcock (2004) concluded that strike-slip basins are relatively small and short-lived, typically in existence for 3–10 Ma. In contrast, basins formed along divergent margins can persist for 100–200 Ma.

Because rhomboidal basins are larger and contain greater sedimentary thicknesses, they are more likely to be preserved in pre-Cenozoic, eroded mountain belts. The now inverted, rhomboidal Stellarton pull-apart basin in the Appalachian Mountains of eastern Canada formed by *c.* 10 km of right-lateral shear during the late Palaeozoic and localized a depocentre now containing up to 3 km of alluvial-fan, coal, and lacustrine sediments (Waldron 2004).

Other areas that contain well-preserved, ancient pull-apart basins include the sheared margins of the Equatorial Atlantic (Benkhelil *et al.* 1995; Miskelly 2002), the southern (Escalona & Mann 2003) and northern (Meneses-Rocha 2001) Caribbean margin; southern California (Crowell & Link 1982; May *et al.* 1993); Mesozoic belts in China (Vincent and Allen 1999; Chen *et al.* 2001; Zhang *et al.* 2003) and Korea (Ryang & Chough 1999); the Mesozoic Iberic range of Spain (Guiraud & Seguret 1985); the Neogene foreland basin of the Betic Range of Spain (Van der Straaten 1993); the Cenozoic Pyrenean Orogeny (Santanach *et al.* 2005); the Mesozoic and Cenozoic basins of the North Sea (Dooley *et al.* 1999); and the Carboniferous orogenic belts of Europe (Mattauer & Matte 1998; Mattern 2001; Shelley & Bossiere 2001; Titus *et al.* 2002).

Reinterpreted pull-apart basins

Subsequent structural and fault kinematic studies by Hossack (1984) have shown that the spectacularly exposed Palaeozoic Norwegian basins are not pull-aparts bounded by steeply-dipping master strike-slip faults as originally proposed on the basis of the original mapping and stratigraphic studies by Steel & Gløppen (1980). Instead, structural studies have shown that the basins formed as rifts above low-angle normal faults related to orogenic collapse along pre-existing, low-angle Caledonian thrust faults. Over 25 km of sediment accumulated in the largest basin over a period of 14 Ma.

The 13.5-km-thick Ridge Basin of southern California was long thought to be a Late Miocene–Pliocene transpressional basin (Crowell

1974a; Crowell & Link 1982), but is now known to be an inverted pull-apart basin formed along a low-angle normal fault imaged by seismic-reflection methods (May *et al.* 1993). The circular Silver Pit of the southern North Sea, that was once interpreted as a pull-apart basin (Smith 2004) is now thought to be a meteorite impact crater (Stewart & Allen 2005).

Pull-aparts in magmatic settings

Pull-aparts have been proposed as important mechanisms for controlling the emplacement of large plutons along trench-linked strike-slip faults in volcanic arc settings such as in the active Central American arc (Girard & van Wyk de Vries 2005), and in ancient orogenic belts (Lacroix *et al.* 1998; Talbot *et al.* 2005).

Complexities and alternative geological models for pull-apart basins

Transform normal extension: an alternative model for pull-apart opening

Ben-Avraham (1992) and Ben-Avraham and Zoback (1992) proposed that there may be a significant component of transform or master fault-normal extension across major strike-slip plate boundary faults like the Dead Sea, San Andreas and Caribbean faults. The observational support for this idea is the large-scale asymmetry seen on 2D seismic-reflection profiles across the basinal fill of the plate-boundary strike-slip faults that, in their view, would require a relatively larger extensional vs. strike-slip component to form. They attributed this basinal asymmetry to fault-normal extension induced by far-field transtensional stress acting on a weak fault. Conversely, far-field transpressional stresses across weak strike-slip faults may lead to a large component of shortening across major strike-slip plate-boundary faults, as observed in well-bore breakouts along the San Andreas fault zone (Mount & Suppe 1987).

Block rotations: an alternative model for pull-apart opening

Detailed seismological studies have shown the existence of rotating fault blocks around vertical axes whose block edges are defined by transverse faults crossing the intervening areas between sub-parallel zones of strike-slip faults. Active examples include the restraining-bend area of the eastern Transverse Ranges (Nicholson *et al.* 1985; Nicholson *et al.* 1986) and the pull-apart area of the Imperial Valley in California (Nicholson *et al.* 1986).

Block rotations may be important to accommodate the progressive passage of one fault block around a releasing bend, as proposed by Neugebauer (1994) for pull-aparts along the North Anatolian fault zone in Turkey. Heimann *et al.* (1990) proposed that large-scale rotation of the basin block of the rhomboidal Barahta pull-apart basin on the northern Dead Sea fault zone provides an alternative model for basin opening (rather than extension along transverse normal faults in the pull-apart model as shown in Fig. 7). Wesnousky (2005) illustrates large-scale block rotations in a 50-km-wide pull-apart segment of the central Walker Lane fault zone in the western USA. Mann (1997) compiled several examples where large-scale plate rotations could lead to transtensional opening and pull-apart formation along major strike-slip faults that bound rotating blocks or plates.

Estimates of rotational strain in pull-apart basins

Waldron (2004) uses the Palaeozoic Stellarton pull-apart basin in Nova Scotia, Canada, to illustrate the significance of rotational strains within the basin. This rotational component of strain is commonly neglected in most structural studies of mature pull-apart basins which accommodate significant offset. For example, at higher shear strains at more advanced stages of pull-apart evolution, previously formed normal faults of the type shown on a simple shear-strain ellipse (e.g. Wilcox *et al.* 1973; Biddle & Christie-Blick 1985) would begin to show inversion along reverse faults and large rotations.

Other approaches and observations on pull-apart basins

Models for pull-aparts based on subsurface geophysical imaging

Due to their thick sedimentary fill and lack of basement exposure, geophysical surveys of pull-aparts, including gravity, refraction and seismic-reflection methods, have improved our understanding of how pull-aparts evolve on now deeply buried master and normal faults. A particular focus of subsurface geophysical surveys in rhomboidal basins has been to determine the exact depth of fill of pull-apart basins and how master faults and transverse normal faults have controlled the observed basement subsidence.

Regional Bouguer gravity mapping by ten Brink *et al.* (1993) and ten Brink *et al.* (1999) of the Dead Sea strike-slip fault has revealed a string of small subsurface basins and strike-slip related depocentres along the trace of the plate-boundary fault

that vary in size, shape, and depth, and bound relatively short (25–50 km) and discontinuous fault segments. These authors suggest that the basins have formed in response to continuous adjustment of the plate boundary to small changes in the relative motion between the African and Arabia plates. Regional aeromagnetic mapping by Hatcher *et al.* (1981) shows how large east–west magnetic anomalies are truncated and offset more than 100 km by the Dead Sea Fault. This is the same amount of offset concluded from several studies of surface bedrock geology.

The depth-to-basement and the dip of basin-controlling normal faults in pull-apart basins are key factors for relating total basin subsidence to offset on master faults (Arbenz 1984; Escalona & Mann 2003; Lazar *et al.* 2006). These studies are also useful for determining whether the sedimentary fill of ancient basins extends deep enough into the crust to be preserved in deeply eroded, ancient mountain belts affected by episodes of widespread strike-slip faulting produced during the final stages of continental convergence.

The depth of penetration of strike-slip faults and associated basins into the crust and upper mantle is controversial because existing geophysical images taken to upper-mantle depths are neither clearly diagnostic nor available for most active and ancient strike-slip fault zones (cf. Lemiszki & Brown 1988; Serpa *et al.* 1988; Rotstein & Bartov 1989; Beaudoin 1994; Garfunkel & Ben-Avraham 1996; Hurwitz *et al.* 2002).

Whether transverse normal faults form low-angle listric or normal faults (Arbenz 1984) or are steeply dipping normal faults as shown in most idealized opening models, awaits improved deep-seismic imaging of the basement surface buried deep beneath pull-apart basins. Using geological mapping and shallow seismicity in the upper crust, previous workers have proposed that overlapping faults bounding releasing and restraining bends merge before reaching mid-crustal levels (Clayton 1966; Sharp 1975; Little 1990; Graymer *et al.* 2007). Others propose that some ancient, exhumed restraining-bend structures can persist to mid-crustal depths (Miller 1994). Deep, wide-angle geophysical profiles of major strike-slip faults indicate that active plate-boundary strike-slip faults may persist as vertical faults through the upper mantle rather than soling out at the base of the crust as predicted by some geological and palaeomagnetic studies (Hole *et al.* 1996; Henstock *et al.* 1997). A more revealing approach has been to use 3D seismic reflection from the petroleum industry to map the internal structure of an ancient Eocene pull-apart in Venezuela (Escalona & Mann 2003).

Confirming the existence of these faults in rhomboidal and extreme basins is key for determining the

value of the pull-apart opening model of Crowell (1974a, b) and Mann *et al.* (1983), where normal faults striking transverse to the master faults are predicted as the primary mechanism for basin opening (Fig. 7). The transform-normal extension of Ben-Avraham (1992) and Ben-Avraham & Zoback (1992) would not require transverse normal faults, and would instead predict basement normal faults parallel to the master faults. Block-rotation models would also not predict regular patterns of transverse normal faults terminating on parallel master faults (Neugebauer 1994). Another issue is the dip of the transverse normal faults: are they listric and low-angle normal faults at right angles to the master faults as proposed by Arbenz (1984) for the Dead Sea pull-apart, or are they more steeply dipping? Geophysical surveys have confirmed the presence of steeply dipping transverse faults consistent with the pull-apart model in the southern basin of the Dead Sea pull-apart (Lazar *et al.* 2006).

Denser grids of seismic-reflection profiles are revealing, in that some basins considered to be rhomboidal pull-aparts on the basis of their surface are in fact underlain by a lazy S- or Z-shaped subsurface basement structure. For example, the northern Gulf of Aqaba (Ehrhardt *et al.* 2005), the Cariaco Basin of Venezuela (Jaimes 2003), and the Sea of Galilee (Hurwitz *et al.* 2002) are underlain by a gentle releasing bend, rather than a rhomboidal pattern of parallel master faults and transverse normal faults as postulated on the basis of earlier seismic-reflection surveys with less resolution and penetration (Schubert 1982; Ehrhardt *et al.* 2005).

Models for pull-aparts at depth, based on high-resolution seismological data

High-resolution recording of earthquakes on creeping strike-slip faults and during large strike-slip events, particularly along the well-monitored San Andreas fault zone, has allowed precise mapping of pull-apart basins down to the base of the brittle crust. Recent monitoring of activity beneath the Cholame pull-apart near Parkfield, California, indicates that the 1-km-wide stepover observed by surficial and shallow fault traces (Shedlock *et al.* 1990) does not extend deeper than 5 km (Bakun *et al.* 2005).

Models for pull-aparts as earthquake rupture boundaries

Pull-apart basins have received considerable attention from seismologists, because of the possibility that they act as boundaries or end-points for the seismogenic propagation of fault slip (Sibson

1985, 1986a, b, 1987). Sibson's hypothesis, based on observed clusterings of aftershocks in the Cholame, Coyote Lake and Mesquite basins on the San Andreas fault system states that ruptures are arrested because of suction forces in the fluid-saturated sediments of the releasing bends. Aftershock clusters following the termination of the mainshock are caused by re-equilibration of fluid pressures in complex dilatational fracture networks. This same process, at deeper levels of the crust, may account for enhanced mineralization in releasing-bend segments (Sibson 1987; Berger 2007; De Paola *et al.* 2007).

Models for pull-aparts based on mesoscopic outcrop studies

Pull-apart structures filled by calcite and quartz are common in brittle faults deforming sedimentary and crystalline rocks. The structural evolution of these outcrop-scale 'pull-aparts' have been described in detail by many workers in different areas where there are surface breaks associated with seismogenic strike-slip faulting (Tchalenko & Ambraseys 1970) and within ancient, deformed rocks (cf. Kim *et al.* 2004, for a recent review).

Despite material property differences and scaling issues, there are many intriguing similarities between these active outcrop-scale structures and sediment-filled pull-apart and fault-termination basins, as described and interpreted by Gamond (1987) and Kim *et al.* (2004). Bahat (1983) and Gamond (1987) propose that sediment-filled pull-apart basins could evolve by the opening along *en échelon* tensile fractures in a manner similar to that seen in outcrop-scale fractures. By the same argument, connecting, compressive 'bridges' are inferred as the manner in which restraining bends form on larger, natural fault systems (Cowgill *et al.* 2004a, b).

Laboratory analogue modelling of pull-apart basins

The simplicity of the natural pattern of pull-apart opening between parallel master faults (Fig. 7) lends itself well to laboratory-scaled sandbox models. Hempton & Neher (1986), Dooley & McClay (1997), Rahe *et al.* (1998), and Basile & Brun (1999) produced analogue models very similar to the main characteristics of pull-aparts known from geological and geophysical studies, including such details as S- and Z-shaped incipient basins, the twin depocentres in rhomboidal basins, and diagonal strike-slip faults marking the extinction of the normal mode of pull-apart basin opening. Sims *et al.* (1999) produced an analogue

pull-apart model in which a weak and strong décollement horizon was introduced at depth. Weak décollements resulted in scattered and asymmetrical sub-basins bounded by oblique-slip faults similar to pull-apart environments like the Gulf of Paria in Trinidad (Babb & Mann 1999; Flinch *et al.* 1999) whilst strong décollements produced more focused basins bounded by normal faults similar to the Death Valley pull-apart.

Despite the close similarities between the geometries of the analogue models and those of the natural examples, all authors modelling pull-aparts state the limitations of applying analogue models to natural systems. For example, these models cannot:

1. accurately simulate the thermal, flexural and isostatic effects generated by or associated with faulting in the upper crust;
2. take into account the effects of pore-fluid pressures and compaction which are to be expected in natural systems (Dooley & McClay 1997); and
3. account for the fact that the modelled fault stepover cannot be bypassed or cut-off at a basement level (T. Dooley, pers. comm. 2006).

However, the modelling by Sims *et al.* (1999) did attempt to address the second problem by inserting a ductile layer designed to simulate buried shale or salt lithologies. Dooley (pers. comm., 2006) notes that in many pull-apart analogue models the presence of a cross-basin strike-slip fault zone indicates that the sidewall fault systems are not 100% efficient and thus some component of differential motion must be accommodated across the basin floor. Despite the presence and activity of the cross-basin strike-slip fault, models show that the stepover continues to expand on its normal faults, no matter how much progressive offset is used for the model.

Numerical models of pull-aparts

Numerical models have been successful in simulating some of the natural features of pull-apart basins, including features of pull-aparts such as the twin depocentres in rhomboidal basins (Rodgers 1980); distribution and orientation of secondary fractures within the master fault overlap area (Segall & Pollard 1980); the symmetry of the basin between the master faults (ten Brink *et al.* 1996); and basin deepening towards the more active master fault (Golke *et al.* 1994; Bertoluzza & Perotti 1997)

Evolution of restraining bends based on geological observations

Introduction

Because active restraining bends form positive topographic features, sometimes reaching several

kilometres in elevation, sites of bends expose older and deeper crustal rocks, with much of their uplift history possibly characterized by an erosional unconformity separating the older rocks in the topographic centre of the bend from surrounding sedimentary rocks at lower elevations that have been eroded off the bend area (Fig. 6a–g). For this reason, techniques used to study restraining bends differ from those described above for pull-apart basins. For example, surface geomorphology and thermochronology of basement rocks are more relevant to restraining-bend uplifts than they are for understanding pull-apart topographic depressions. Conversely, seismic-reflection imaging has had little impact on understanding restraining bend uplifts—particularly if the uplifts are composed of non-layered and difficult-to-image crystalline rocks.

Because of these limitations and the fact that pull-apart basins are far more numerous than restraining bends (cf. compilations by Aydin & Nur 1982; Bahat 1983; Table 2 of this paper), there has been significantly less of a global effort in the classification, mapping and modelling of restraining bends than there has been for pull-apart basins. Unlike pull-apart basins, the morphological classes of restraining bends shown on Figure 1b occur on very different scales, and in some cases, in different tectonic settings. For this reason, there is presently no observationally based, geological continuum model – as in the case of pull-apart basins (Fig. 7).

In this review of restraining bends, I will review existing classification schemes and propose a new restraining-bend classification (Fig. 1c) based on the global compilation of restraining bends on Table 2. It is hoped that this preliminary descriptive scheme for classifying restraining bends will lead future workers to derive a more comprehensive and genetic classification.

Morphological classification of strike-slip restraining bends

Classification

Using previous terminology proposed by Crowell (1974a, b) and Mann & Gordon (1996), I propose a threefold morphological classification of active and ancient restraining bends, shown in Figure 1c. The three bend classes include:

Transpressional uplifts. These are adjacent to straight segments of strike-slip faults that are oblique to the direction of plate or block motion (Fig. 1c). Technically, this class is not a fault ‘bend’, since the main strike-slip fault remains straight or very gently curving. Nevertheless, I

include this type in the classification because a narrow, elongate topographic uplift and exposure of older rocks is produced as the result of the oblique intersection of the fault with the direction of plate/block motion (Fig. 1c). These uplifts can persist for along-strike distances of tens to hundreds of kilometres, and they combine elements of strike-slip (simple shear) with an element of shortening (pure shear). Examples of transpressional uplifts include the uplift of narrow strips of oceanic crust, including ophiolites along the Macquarie Ridge south of New Zealand (Massell *et al.* 2000; Fig. 6a); narrow strips of mid- to deep levels of continental crust along the Alpine Fault of New Zealand (Little *et al.* 2005); the Border Ranges fault zone of SE Alaska (Little 1990; Roeske *et al.* 2003); and the Betic Range of SE Spain (Keller *et al.* 1995).

Much of the 1000-km length of the two continuous faults forming straight segments of the Macquarie Ridge transpressional uplift are slightly oblique (c. 5–10°) to the direction of Australia–Pacific plate motion, and for that reason control the uplift and exhumation of narrow (10–40 km) transpressional uplifts of sheared oceanic basement rocks adjacent to the active strike-slip fault zones (Massell *et al.* 2000; Meckel *et al.* 2005; Fig. 6a). Increasing obliquity of plate motion in a southward direction leads to widening of the elongate transpressional uplift as seen on the southern segment of the Macquarie Ridge that exposes a partial ophiolite sequence on Macquarie Island (Daczko *et al.* 2003; Fig. 6a). Within this broad transpressional system, smaller pull-aparts coexist along the main trend of the strike-slip fault that bisects the transpressional ridge (Daczko *et al.* 2003; Fig. 6a).

Sharp restraining bends. These form at discrete fault stepovers and produce rhomboidal topographic uplift and exposure of older rocks (Fig. 6b & c). Sharp restraining bends are generally much smaller in size (fault separations generally <15 km) than the third and most prevalent type of bend, the *gentle restraining bend* (fault separations <200 km; Fig. 6d–g; Table 2). Two examples of mapped, sharp restraining bends deforming Late Neogene, clastic sedimentary rocks along active strike-slip faults are the Villa Vasquez bend (length 14 km, width 2 km) along the Septentrional fault zone in Hispaniola (Mann *et al.* 1999; Fig. 6b) and the Ocotillo badlands bend along the Coyote Creek fault segment of the San Jacinto fault zone in southern California (width 1.5 km, length 8 km) (Sharp & Clark 1972; Brown *et al.* 1991; Lutz *et al.* 2006; Fig. 6c).

The relatively small sizes and rarity of active sharp restraining bends probably indicate that they are easily bypassed, become extinct, and are

eroded during progressive offset of the strike-slip fault zone. For example, dating and palaeomagnetic work by Brown *et al.* (1989) has shown that the sharp restraining bend forming the uplifted topography of the Ocotillo badlands formed less than 1 Ma; only reflects a small fraction of the total offset on its master faults; and is accompanied by a limited amount of vertical-axis rotation (Fig. 6c). As in the case of pull-apart basins, formation of sharp bends may be linked to small lateral shifts or ‘switchyard behaviour’ in the strike-slip fault trace (Brown *et al.* 1991) – perhaps in response to a changing pole of rotation (Fig. 9).

Wesnowsky (1988) has compiled the fault step-over distances for both pull-apart and restraining bends on the San Andreas and North Anatolian fault zones, and concluded that faults with greater total offsets exhibit fewer restraining and releasing bends and overall straighter fault traces. This observation confirms the idea that smaller irregularities like sharp bends with fault separations less than a few kilometres are bypassed or ‘smoothed’ by faults with greater offsets, and therefore do not evolve into longer sharp restraining bends. Earthquake rupture data compiled by Harris *et al.* (2002) and Graymer *et al.* (2007) show that bends less than 5 km wide generally do not act as rupture boundaries for ground breaks for strike-slip earthquakes like the 1999 Izmit earthquake in Turkey.

Gentle restraining bends. These form the familiar shapes colloquially called *lazy-S shape* for gentle bends between right-lateral faults, as shown in Figure 1C, and *lazy-Z shape* for those between left-lateral faults. As with evolving pull-apart basins (Fig. 7), the S and Z shapes of gentle restraining bends in strike-slip plate boundary settings (Fig. 2) become more prominent as the master-fault separation ranges upward from values of 1.3 km (Echo Hills restraining bend, Nevada – Campagna & Aydin 1991) to values of 6 km (Montejunto restraining bend, Portugal – Curtis 1999), 20 km (Santa Cruz restraining bend, California, Burgmann *et al.* 1994, Fig. 6d) to greater than 20 km (San Bernardino restraining bend, eastern Transverse Ranges, California, Spotila *et al.* 2001, Fig. 6e).

These gentle, bend fault geometries produce elongate to domal topographic uplifts mini-fold–thrust belts; localized exposure of older deeper crustal igneous and metamorphic rocks in the core of the uplift; and a radiating pattern of alluvial fans and fan deltas that accommodate the erosional stripping of the bend uplift (Mann & Gordon 1996; Fig. 6d–g). The lineaments marking the master strike-slip fault traces in the bend area itself are sometimes obscured by the rugged topography of the core of the bend, but are well expressed by valleys of the master faults on either side of the

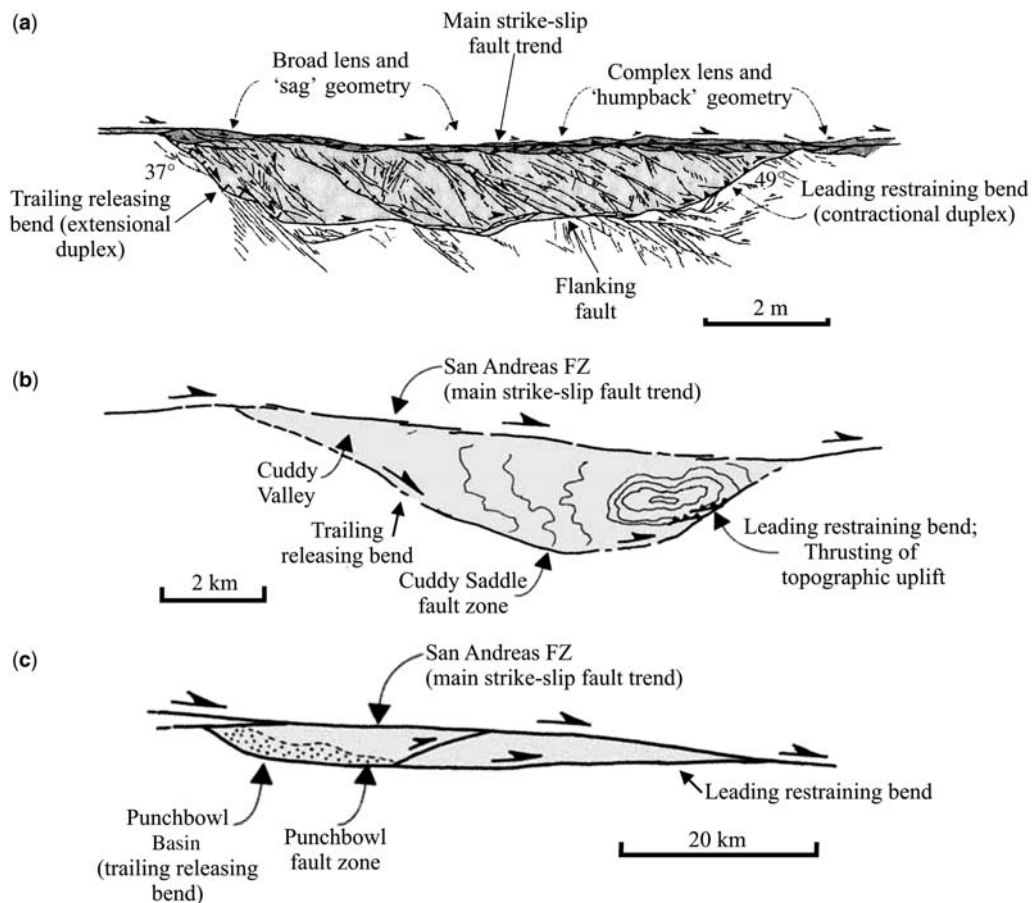


Fig. 10. Swanson (2005) proposes that 'sidewall ripouts' are scale-independent, based on his compilation of fault geometries for active and ancient strike-slip faults. (a) Outcrop-scale (12-metre-long) example of a 'sidewall ripout' (light grey) mapped from a Palaeozoic outcrop at Cape Elizabeth, Maine, by Swanson (2005). This structure exhibits imbrication and thrusting in a single obliquely thrusting 'leading ramp', fault and extension along three normal, oblique-slip faults in a 'trailing ramp'. Darker grey indicates the complexly deformed and linear, main strike-slip fault zone. (b) Regional-scale (10-km-long) example of a Neogene 'sidewall ripout', compiled by Swanson (2005) from mapping of the San Andreas and Cuddy Saddle fault zones at the western end of the Transverse Ranges by T. Davis and Duebendorfer (1987). Area of thrusting and topographic uplift is interpreted as a 'leading ramp'; the low-lying Cuddy Valley is interpreted as a 'trailing ramp'. (c) Regional-scale (55-km-long) example of a Neogene 'sidewall ripout', compiled by Swanson (2005) from mapping of the San Andreas and Punchbowl fault zones by Dibblee (1968). The Punchbowl Basin was filled by Neogene clastic sedimentary rocks, and is interpreted as a 'trailing ramp'.

bend (e.g. Massif du Macaya restraining bend in Haiti, Mann *et al.* 1995).

Gentle restraining bends are by far the most common, as compiled for this paper, and are the largest in their areal extent. Two examples of lazy-S gentle restraining bends along the San Andreas fault system are shown on Figure 6d & e. The Santa Cruz restraining bend on the San Andreas fault zone in northern California formed by a fairly subtle 5–10° change in the strike of the San Andreas fault zone (Anderson 1990, 1994; Schwartz *et al.* 1990; Burgmann *et al.* 1994;

Fig. 6d). The San Bernardino Mountains bend on the San Bernardino fault zone of southern California forms a more prominent lazy-S shape created by a larger 15–20° change in strike of the southern San Andreas fault zone (Spotila *et al.* 2001; Fig. 6e).

Both gentle bends exhibit fault separations seven to ten times the length of fault separations for the smaller sharp restraining bends shown in Figure 6b & c. Since these gentle bends have larger fault separations, bypassing with increasing fault offset (Wesnousky 1988) would be less

likely, and therefore the bends might be expected to persist longer in the rock record. Using thermochronological methods, Burgmann *et al.* (1994) proposed that the Santa Cruz bend was originally uplifted at 12 Ma; Spotila documents at least a 1.8 Ma uplift history for the San Bernardino bend using mapping and thermochronology; Blythe *et al.* (2000) document the beginning of the bend-related uplift of the San Gabriel Mountains to have started at 7 Ma.

The gentle restraining bends with the largest areal coverage are found in a particular active tectonic environment: subduction to strike-slip transition areas (Bilich *et al.* 2001). These areas are the tectonic transitions between strike-slip plate boundaries and subduction boundaries, and include the gently curving boundaries between the North America and Caribbean plates in northern Central America (Mann & Gordon 1996; Fig. 6f); the North America and Caribbean plates in the Greater Antilles (Mann *et al.* 2003; Fig. 6g); and the South America and Scotia plates in southernmost South America (Cunningham 1995). The curvature of the faults near their intersection is related to either trenchward propagation of the strike-slip faults as described by Govers & Wortel (2005) or oblique collisional indentation as produced by the Bahama carbonate platform on the NE Caribbean plate (Mann *et al.* 2003; Fig. 6g).

Pattern of adjacent restraining and releasing bends at outcrop to regional scales

Sidewall ripouts on mesoscopic strike-slip faults

Swanson (1989a) reported asymmetrical sidewall ripout geometries from coastal exposures of mesoscopic, Late Palaeozoic strike-slip fault zones in southern Maine and western Greenland (Fig. 10a). *Sidewall ripouts* in plan view are plano-convex fault lenses and slabs where a splay fault diverges away from a section of the main fault to create an oblique-slip *flanking fault*, which rejoins the main trace on another lateral ramp further along strike (Fig. 10a).

The overall fault pattern formed in plan view is an asymmetrical fault lens flanking the main strike-slip fault trace. The asymmetry of the fault lens leads to the development of a gentle restraining bend or *contractional duplex* (Woodcock & Fischer 1986) at the leading edge of the fault lens, and the development of a gentle releasing bend or *extensional duplex* at the trailing edge of the fault lens (Swanson 1989b; Fig. 10a). Slight bulging or

'humpback' geometry deforms the straight fault trace to accommodate displacement at the restraining bend, whilst 'sag' geometry accommodates displacement at the releasing bend.

Swanson (1989a) interpreted the kinematic mechanism for the formation of mesoscopic ripouts to be the result of adhesion and increased friction on the main fault trace, followed by a lateral jump of the active fault surface to the adjacent curving fault trace and activation of the adjacent restraining and releasing bends. Revived movement on the locked, straight fault trace may lead to deactivation of the flanking fault and the shearing off and abandonment of the entire fault lens.

Significance of sidewall ripouts at regional scales

Swanson (2005) recognized similar sidewall ripout fault patterns on many active crustal-scale, strike-slip faults in all three types of strike-slip faults worldwide (Fig. 2). Proposed examples of regional-scale ripouts on *boundary strike-slip faults* included the San Andreas and San Gabriel fault zones of southern California (Fig. 10b & c). His proposed examples of *trench-linked strike-slip faults* with sidewall ripouts with adjacent releasing and restraining bends included the Atacama and Domeyko fault zones in Chile and the Yalakom fault zone of British Columbia. His proposed examples of *indent-linked strike-slip faults* included the Anninghe, East Xiaojiang, and Qujiang fault zones of the SE Tibetan Plateau in China.

For active crustal-scale faults, the surface expression of ripout translation includes a *leading gentle restraining bend* characterized by high topography, erosion, folding and oblique thrusting, and a *trailing gentle releasing bend* characterized by low topography, normal faulting and clastic sedimentation (Swanson 2005) (Fig. 10). Active, crustal-scale ripout slabs identified by Swanson (2005) range from 5 to 500 km in length. For example, his proposed 15-km-long ripout defined by the straight San Andreas Fault and the arcuate Cuddy Saddle Fault at the western end of the Transverse Ranges in southern California, shows a low valley topography at its trailing end of the proposed ripout structure and an uplifted topography with minor thrusting at its leading edge (Davis & Duebendorfer 1987; Fig. 10b). A larger, 65-km-long, 0.5-km-wide ripout structure is defined by the straight San Andreas fault zone and the arcuate Punchbowl fault zone (Dibblee 1968; Fig. 10c). The trailing edge of the ripout is defined by the Punchbowl Basin, filled by Neogene clastic sedimentary rocks (Dibblee 1968).

Paired strike-slip fault bends

Several previous workers have noticed the tendency for restraining and releasing bends to occur in pairs adjacent to one another and to exhibit almost the identical amount of master-fault separation (Figs 1d & 9a). In other words, the net effect of a paired bend is to locally deviate the main strike-slip fault trace, but to return the trace to the original and regional strike of the fault once the bend is passed (Fig. 1d). In map view this is exactly the same fault pattern as shown for regional ripout structures proposed by Swanson (2005; Fig. 10a–c). However, an alternative term for this pattern of adjacent restraining and releasing bends that I will use in this discussion is *paired bends*, which, unlike the term *sidewall ripouts*, is purely descriptive and has no genetic connotations of tectonic origin.

In Figure 11, I show four examples of paired bends that are described in greater detail in the latter part of this paper, they range in master-fault separations from 3 km in length for the Coyote Creek fault zone in southern California (Fig. 11a), to 50 km for the North Anatolian fault zone in the Sea of Marmara (Fig. 11b), to 175 km for the San Andreas fault system of southern California (Fig. 11c), to 50 km along the Dead Sea fault system (Fig. 11d). Each of these examples shows a similar pattern of an array of *en échelon* pull-apart basins adjacent to a large gentle restraining bend. Following the work of Swanson (2005), it became apparent to me that the paired bend, or crustal-scale ‘sidewall ripout’, is present at many scales within the plate boundary and trench-linked settings for strike-slip faults (Woodcock 1986; Fig. 2 & Table 1).

Comparison of tectonic models for restraining bends

Regional tectonic controls to bend fault traces

The same regional tectonic mechanisms summarized in Figure 8 for bending strike-slip fault traces to form pull-apart basins may also apply to natural examples of individual restraining bends (Figs 6a–g) and to paired bends (Fig. 11a–d). However, in this section, I will compare the predictions of three additional geological mechanisms with those shown in Figure 6 that have been proposed specifically for individual restraining bends and paired bends.

Shear-zone model

Gamond (1987) and Cowgill *et al.* (2004a) inferred that the origin of single and paired restraining bends

at outcrop to regional scales was a natural consequence of progressive evolution and offset in the strike-slip fault zone (Fig. 12a). In this model, two sets of *en échelon* features form as an initial configuration in a developing shear zone: first a diffuse set of R-shears that eventually link up via bridging P-shears in an evolutionary sequence that has been observed and well documented in analogue models using:

1. various laboratory media (Tchalenko & Ambraseys 1970; Hempton & Neher 1986);
2. surface ruptures of natural strike-slip faults deforming relatively isotropic Quaternary sediments (Tchalenko & Ambraseys 1970); and
3. and in mesoscopic outcrops of rocks that have experienced brittle deformation (Gamond 1987).

The result of the shear-zone evolution is the present configuration consisting of a through-going fault trace with a sinusoidally curving trace that would form the sites of paired bends during progressive shear. Another model for the formation of a large single restraining bend mapped by Cowgill *et al.* (2004a, b) along the Altyn Tagh Fault includes rotation of large crustal blocks that deformed the fault zone and formed stepped bends in the fault as the fault propagated from west to east.

Sidewall ripouts

Swanson (1989a) interpreted the kinematic mechanism for the formation of mesoscopic ripouts at mesoscopic (Fig. 10a) and crustal (Fig. 10b & c) scales to be the result of adhesion or increased friction on the main strike-slip fault trace, followed by a lateral jump of the active fault surface to the adjacent curving fault trace and activation of the adjacent restraining and releasing bends (Fig. 12b). This model is similar to the finite-element model of Bridwell (1975) for forming sinusoidal strike-slip fault traces.

Revived movement on the locked, straight fault trace may lead to deactivation of the flanking fault and the shearing off of the entire asymmetrically shaped fault lens or ‘sidewall ripout’ (Fig. 12b). In his crustal-scale examples from the San Andreas and other fault systems, Swanson (2005) used dating of faulting and basinal deposits to show the history of abandonment and reactivation of the main fault trace and the splay fault defining the sidewall ripout (Fig. 10b & c).

Sidewall ripouts along strike-slip faults are analogous to lenses or ‘horses’ in contractional duplex structures on thrust faults. Progressive thrust offset leads to ‘shunting’ of the horse from the hanging wall to the footwall and then back to the hanging wall (Woodcock & Fischer 1986) (Fig. 12b).

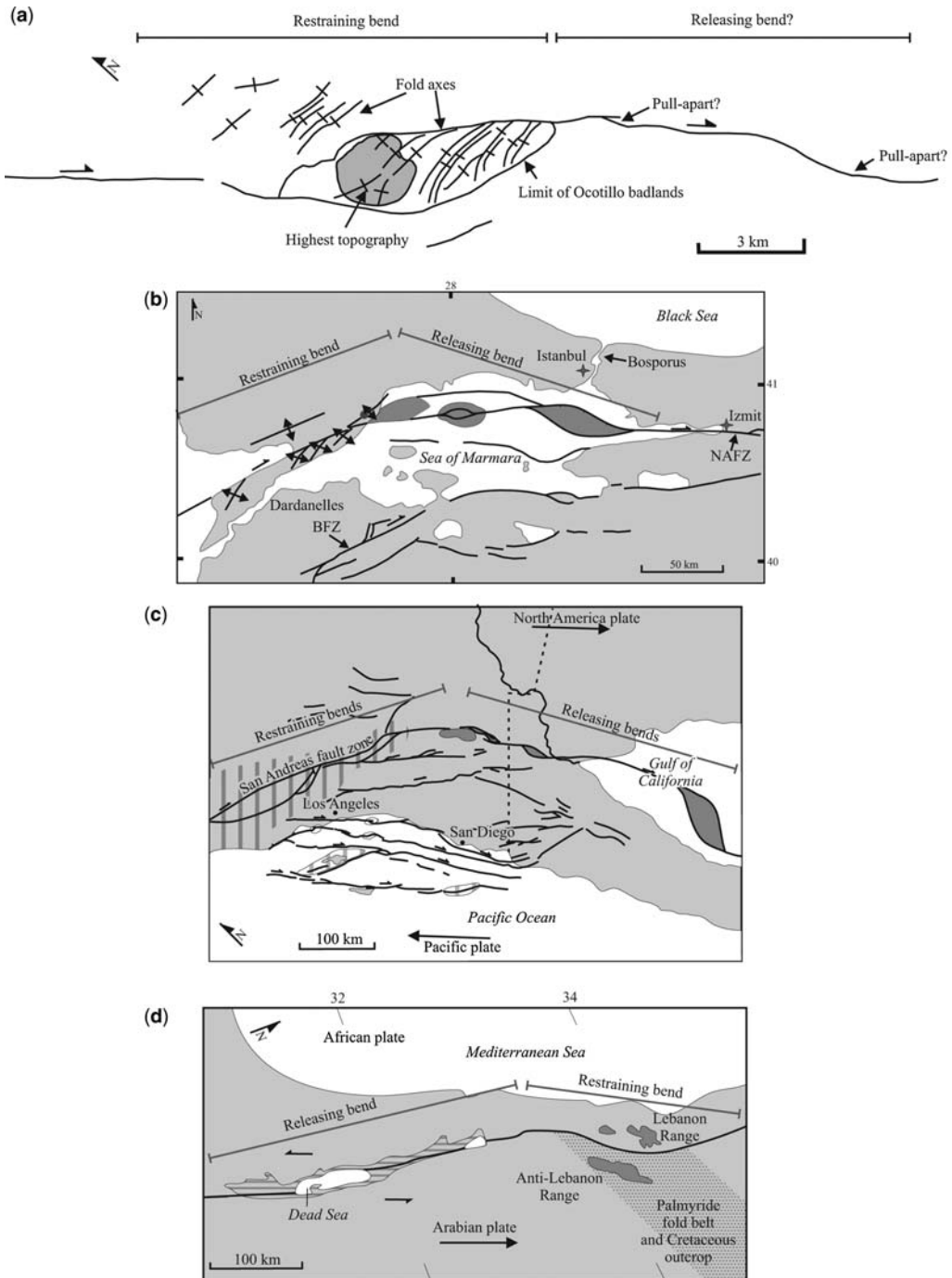


Fig. 11. Examples of large-scale paired-bend structures. The main strike-slip fault appears to contain releasing bends in one direction of the fault and restraining bends in the opposite direction. (a) Schematic map of the Ocotillo restraining bend along the Coyote Creek fault zone in southern California, modified from Segall & Pollard (1980), based on the original mapping of Sharp & Clark (1972). An inferred paired restraining- and releasing bend is indicated. (b) Sea of Marmara paired bend. (c) Southern California paired bend. (d) Dead Sea–Lebanon paired bend.

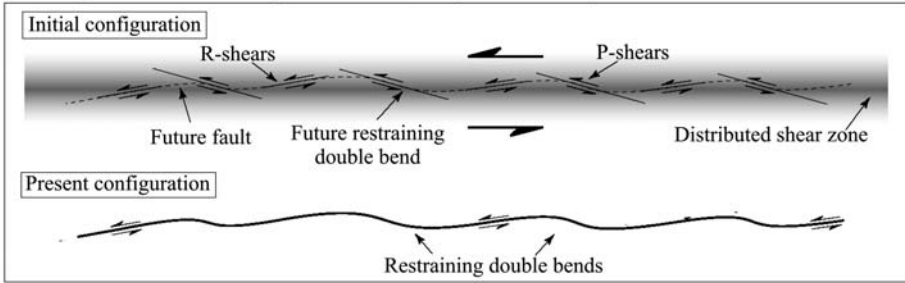
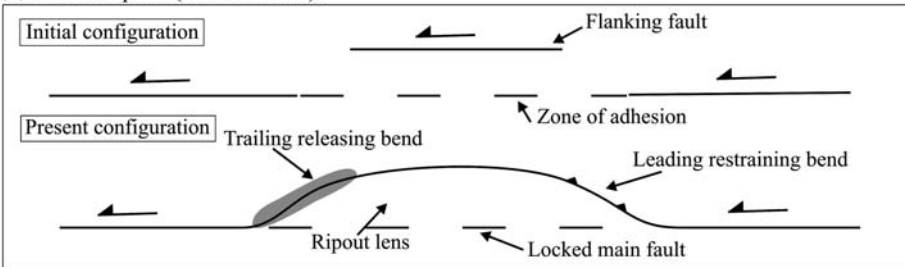
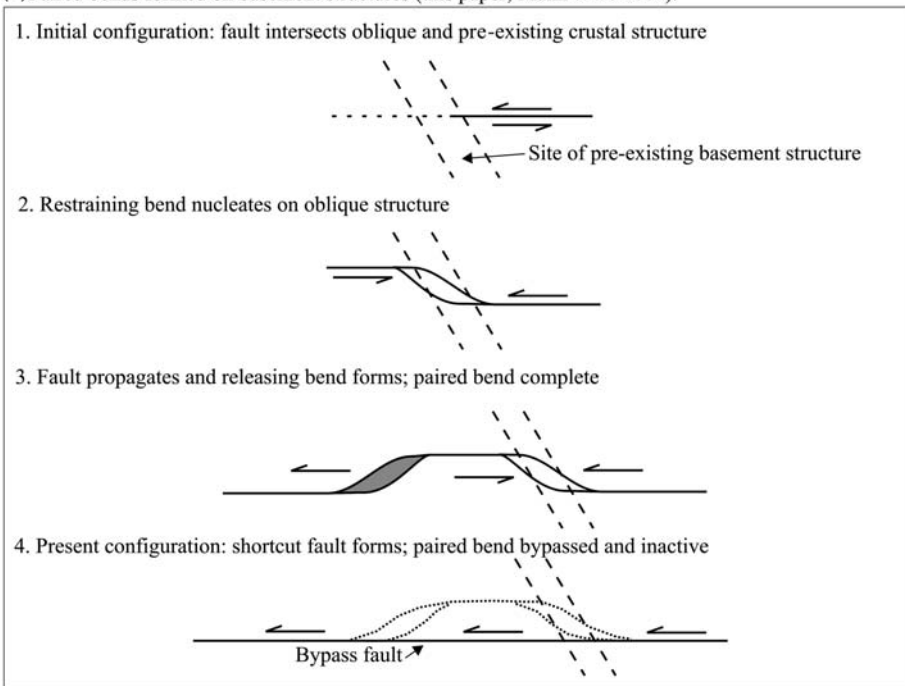
(a) Linkage of P- and R-shears (Gamond 1987; Cowgill *et al.* 2004b).**(b)** Sidewall ripouts (Swanson 2005).**(c)** Paired bends formed on basement structures (this paper; Mann *et al.* 2007).

Fig. 12. (a) Model for the formation of adjacent releasing and restraining bends (i.e. double bends or paired bends) along strike-slip faults, based on the linkage of P- and R-shears (modified from fig. 12 of Cowgill *et al.* 2004b; see also Gamond 1987). (b) Model for the formation of adjacent extensional and contractional 'ramps' at the edges of 'sidewall ripouts' based on mesoscopic mapping and compilations of regional strike-slip fault traces, basins, and bends by Swanson (2005). (c) Model for the formation of adjacent releasing and restraining 'paired bends', based on Mann *et al.* (2007a) and the compilation done for this paper.

Basement control on paired bends

Mann *et al.* (2007a) propose a model based on the Jamaican restraining bend for the interaction of a propagating strike-slip fault and a pre-existing basement structure (Fig. 12c). In the Jamaican example, the pre-existing structure is a Palaeogene rift that trends at a 45° angle to the strike-slip zone and precedes the onset of strike-slip faulting. One puzzling question is: why does the strike-slip fault return to its original trend by forming an adjacent pull-apart basin as shown in the third drawing in Figure 12c? An identical map-view fault pattern is seen along other paired bends, as shown on the four paired-bend examples compiled in Figure 11.

One structural explanation is that the pull-apart basin forms as a geometric accommodation to shortening at the restraining bend localized on the pre-existing rift structure (Woodcock & Fischer 1986). To maintain zero bulk-volume change in a plane-strain strike-slip system, the excess volume of deformed crust 'piling up' at the restraining bend during progressive strike-slip displacement must be balanced by a volume deficit by thinning the crust in the adjacent, along-strike area of the releasing bend. Since mass balance of the crust is preserved, the area of thinned crust is equivalent to the thickening in the restraining bend, and therefore the approximate shapes and master-fault separations would remain similar. Whilst some of these crustal excesses or deficiencies will be accommodated by localized uplift and subsidence of the ground surface near both bends, most of the crustal deformation at the bend will be propagated laterally through the crust – hence the need for a releasing bend to develop adjacent to the restraining bend, and for the master fault to shift laterally to its original trend, as shown schematically in Figure 1d and in the natural examples of sidewall ripouts/paired bends in Figures 10 & 11.

Different approaches and observations on natural restraining bends

Tectonic geomorphology

A prominent feature of strike-slip restraining bends is their localized steep topography that commonly forms anomalously high peaks compared with the surrounding region (Fig. 6a–g). The steep drainage slopes and high local relief of the elongate Santa Cruz Mountains are strongly symmetrical about a lazy-S-shaped left-stepping restraining bend in the San Andreas fault zone (Fig. 6d), and this geomorphology has been modelled to support its tectonic origin by advection of crust and shortening around the bend (Anderson 1990, 1994; Burgmann *et al.* 1994). Schwartz *et al.* (1990) note that horizontal

compression across the bend is consistent with the large thrust component of the 1989 Loma Prieta earthquake; the abundant folds and faults mapped in the southern Santa Cruz Mountains by Aydin & Page (1984); and the steep SW dip of the San Andreas fault zone beneath the mountains.

Gomez *et al.* (2006, 2007) used a digital elevation model to map high-altitude, low-relief surfaces in the 200-km-long Lebanese restraining bend. These surfaces are used as reference points to assess net post-Miocene uplift and strain partitioning effects.

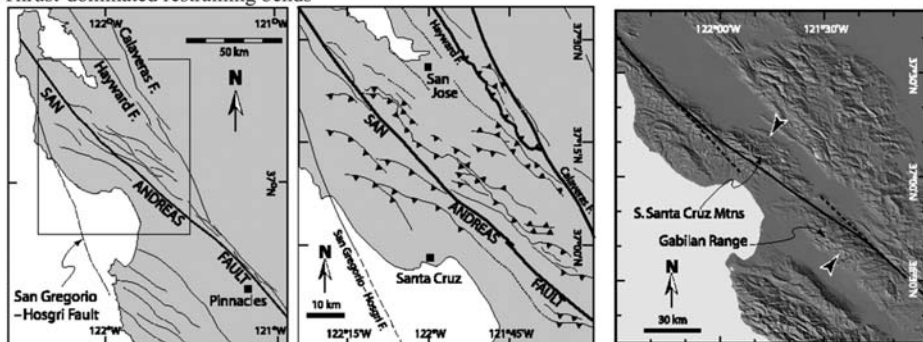
Cowgill *et al.* (2004a) use a digital elevation model to document the tectonic geomorphology of the Akato Tagh gentle bend along the Altyn Tagh Fault in Tibet. Topographic profiles show that the bend uplift is asymmetrical, with the highest and widest areas of the range expressed as two topographic maxima located within the inside corners of the double bend. Structural studies show that the topographic elevated inside corners of the bend are regions of strong fault-perpendicular shortening, and that this shortening may cause the master fault to rotate and reduce and 'smooth' the restraining-bend angle through time (Wesnousky 1988).

Cunningham (2007) reviews the tectonic geomorphology of 12 individual restraining bends and fault-termination zones in the Altai, Gobi Altai and Tien Shan of central Asia. Cretaceous–Palaeocene summit peneplains provide a useful topographic datum for determining relative amounts of topographic uplift and the structural asymmetry of the ranges.

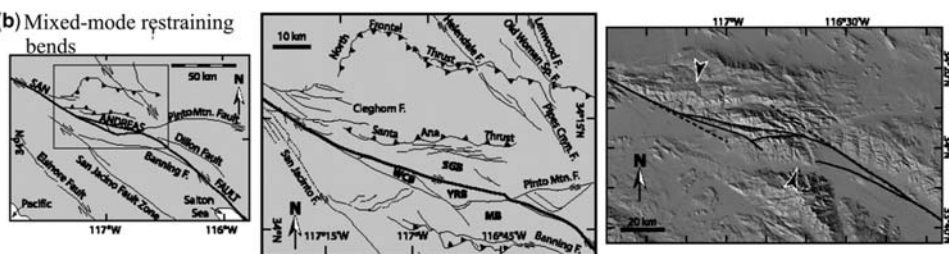
Comparative structure and tectonics of major restraining bends

The regional structure and seismicity of the larger and better-studied gentle restraining bends, including the San Gabriel and San Bernardino Mountains of the Transverse Ranges of California; the Lebanese bend of the Dead Sea Fault; the Hispaniola bend of the northern Caribbean; and the Southern Alps of New Zealand (reclassified in this paper as a transpressional ridge – Fig. 1b) have been compared at regional scales by Scholz (1977), Mann *et al.* (1984), Mann & Burke (1984), Yeats & Berryman (1987), and Cowgill *et al.* (2004a). On the basis of their mapped structure, Cowgill *et al.* (2004a) classify the Santa Cruz gentle bend (or double bend in their terminology) as a *thrust-dominated bend* with the highest elevations, steepest slopes, and largest local relief, all within 10 km of the San Andreas Fault and bounded by inside topographic corners of the bend in the Santa Cruz Mountains and Gabilan Range (Burgmann *et al.* 1994) (inside corner highs are indicated by arrows in Fig. 13a).

(a) Thrust-dominated restraining bends



(b) Mixed-mode restraining bends



(c) Strike-slip-dominated restraining bends

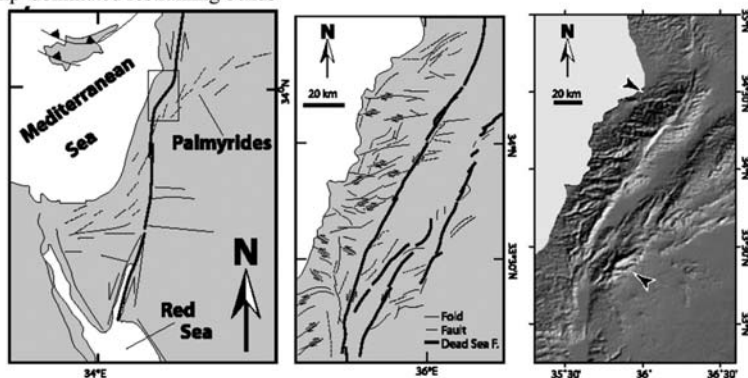


Fig. 13. Three types of restraining bends based on Cowgill *et al.* (2004b). (a) Thrust-dominated Santa Cruz Mountains. (b) Mixed-mode Big Bend on the San Andreas Fault in southern California. (c) Strike-slip-dominated Lebanon Ranges. Black triangles indicate corner topographic highs.

The San Bernardino Mountains exhibit similar nodes of topographic uplift in the inside corners, but their structure is classified by Cowgill *et al.* (2004b) as a *mixed-mode bend* by combining thrusting in the outside of the bend and strike-slip faulting in its internal areas (Fig. 13b). The Lebanese bend, which exhibits similar inside corner topographic uplifts, is interpreted as a *strike-slip dominated bend*, although Gomez *et al.* (2007) describe a mixed-mode, or strain-partitioned, style of deformation with internal strike-slip faulting and an external zone of folding and thrusting (Fig. 13c).

Calais *et al.* (2002) and Mann *et al.* (2003) proposed a similar mixed-mode, or strain-partitioned, style of deformation for the Hispaniola bend, using GPS results (Fig. 6g).

McKenzie (1972) first noted that triangular-shaped crustal wedges in the vicinity of the Transverse Ranges bend of southern California may indicate a process of 'tectonic escape' or lateral migration of continental blocks away from the maximum area of bend-related shortening in the central Transverse Ranges. This same idea of tectonic escape was proposed in a more recent paper by

Walls *et al.* (1998) that summarized geological and earthquake evidence for east–west crustal escape along strike-slip and oblique-slip faults and north–south shortening by thrusting and conjugate strike-slip faulting in the region of the Transverse Ranges. This escape style of tectonics could be classified as ‘mixed mode’ in the terminology of Cowgill *et al.* (2004a), since the kinematics involve a shortening and strike-slip escape component.

Geophysical imaging and the deep structure of restraining bends

Because bends form positive, mountainous areas, they are not conducive for acquiring seismic-reflection data. For this reason, the deep structure of bends remains poorly known and controversial. For example, R. S. Yeats has argued both for shallow detachment of the Transverse Ranges as a ‘crustal flake’ (Yeats 1981) and for its deeply rooted structure in an undetached thick crust (Yeats *et al.* 1994). Balanced structural cross-sections drawn by Namson and Davis (1988) for the western Transverse Ranges invoke a major inferred detachment in the mid-crustal region.

Seismic-reflection data show that the New Zealand transpressional uplift is a pop-up-type structure on 2D seismic profiles that extends to depths of 20–30 km (Little *et al.* 2005). Industry seismic-reflection data from Syria show that the Lebanese restraining bend formed on a pre-existing Cretaceous rift system now inverted to form a 100-km-wide fold–thrust belt (McBride *et al.* 1990). Industry seismic-reflection data from the Hispaniola restraining bend show a ‘mixed-mode’ style of strain partitioning, with outer fold–thrust belts and subduction zones flanking an interior zone of strike-slip faults (Mann *et al.* 2003).

Seismological data to map downward continuation of restraining bends

High-resolution recording of earthquakes on creeping strike-slip faults and during large strike-slip events – particularly along the well-instrumented and monitored San Andreas fault zone has allowed precise mapping of restraining bends down to the base of the brittle crust. For example, Schwartz *et al.* (1990) have shown that aftershock hypocentres of the 1989 Loma Prieta earthquake define a 5-km-wide, 10-km-deep, SW-dipping restraining bend between the San Andreas and Sargent–Berrocal fault zones. The Mission Hills restraining bend connects the Hayward and Calaveras faults and forms a topographic uplift and one of the fastest uplifting areas in this region (based on radar interferometry, Burgmann *et al.* 2006). Manaker *et al.* (2005) have used relocated earthquakes to define the bend down to a depth of 10 km.

Graymer *et al.* (2007) review evidence for other down-dip extensions of other small restraining bends along the San Andreas Fault, determined using high-resolution earthquake hypocentres. They conclude that restraining or releasing bends with master-fault separations of less than 5 km are often not expressed as two faults in the subsurface. This result is consistent with the finding of Harris *et al.* (2002) that the Izmit (Turkey) strike-slip earthquake of 1999 ruptured through bends with fault separations of less than 5 km and fault angles of less than 30°.

Deep crustal bends and preservation potential

The depth of penetration of restraining bends into the mid- and lower crust is a key question for understanding their preservation potential in deeply eroded ancient mountain belts (Little 1990). Miller (1994) mapped an example of a mid-crustal example of a rhomboidal restraining bend with a 10 km fault separation in the Cascade Mountains of Washington State, and concluded that an upper crustal stepover penetrated to the mid-crust. Corsini *et al.* (1996) describe a ductile, mid-crustal restraining bend formed in the Brazilian craton during the Late Precambrian. The deformation pattern is very similar to ramp–flat transitions in thrust tectonics (Woodcock & Fischer 1986).

Palaeomagnetic data from restraining bends

Palaeomagnetism has been used to study large-scale block rotations about vertical axes within the gentle Transverse Ranges restraining bend by Luyendyk (1991; Fig. 13a); widespread vertical-axis block rotations within the gentle Lebanese restraining bend (Ron *et al.* 1990; Fig. 13c); and vertical axis rotations within smaller rhomboidal-type bends including the Ocotillo badlands (Brown *et al.* 1991; Fig. 11a) and the rhomboidal Almacik restraining bend or inferred ‘crustal flake’ actively being displaced and rotated along the North Anatolian fault zone (Saribudak *et al.* 1990; Table 2).

Uplift history and thermochronology of restraining bends

Fission-track dating has been used to study the localized uplift history of the Denali restraining bend in Alaska (Fitzgerald *et al.* 1993), the Transverse Ranges bend (Blythe *et al.* 2000; Spotila *et al.* 2001) and the Santa Cruz mountains bend (Burgmann *et al.* 1994). Spotila *et al.* (2001) note that the uplift mechanism of deep-crustal rocks is most

likely a response to the gentle bend fault geometry rather than larger-scale kinematic models calling for fault-normal shortening along the entire length of the San Andreas fault zone (Mount & Suppe 1987). Cowgill *et al.* (2004a) noted that the anomalous topographic elevations marking one inside corner of the Santa Cruz gentle bend (Fig. 13a) correspond with one area of rapid uplift as measured by fission-track dating (Burgmann *et al.* 1994).

Modelling approaches to understanding restraining bends

Block and kinematic models

Several areas of large-scale (30°), counterclockwise block rotations have been identified in the Lebanese restraining bend (Ron *et al.* 1990). These rotations have been modelled either as neotectonic block rotations associated with the present-day bend (Ron 1987; Westaway 1995) or an earlier deformation related to the inversion of a Cretaceous rift system and the formation of the Palmyride fold–thrust belt (Walley 1998; Westaway 2004). Gomez *et al.* (2007) attempts to reconcile both interpretations by proposing a poorly dated two-stage history:

1. an earlier deformation involving less-oblique plate motion at the bend and controlled by a more distant pole of rotation; this deformation phase was characterized by distributed strike-slip faulting, less crustal shortening and large-scale block rotations (Ron 1987); and
2. a later phase of deformation involving increased oblique plate motion and increased strain partitioning into an inner zone of strike-slip faulting and an outer zone of folding and thrusting.

The palaeomagnetic recognition of significant vertical axis rotations in the region of the Transverse Ranges of southern California has led to block models by Luyendyk (1991), Dickinson (1996), and Ingersoll & Rumelhart (1999). All these models propose large rigid blocks that experience greatest rotations in the restraining-bend area in the Miocene and Pliocene, and that basinal areas were created at their block edges. GPS surveys by Feigl *et al.* (1993) show only smoothly varying motion of crustal material around the Big Bend area of southern California, and therefore indicate that all large-scale block rotations have ended.

Analogue models of restraining bends

McClay & Bonora (2001) have produced laboratory analogue models of restraining bends that simulate many of the observed features of natural bends. Three experimental conditions are produced:

underlapping master faults which produce ‘lozenges’ or gentle bends; master faults with no overlap that produce rhomboidal basins; and master faults with overlap that also produce rhomboidal basins. In cross-sections, the bends show pronounced structural asymmetry and doubly plunging anticlines. As in analogue models of releasing bends, the models cannot accurately simulate the anisotropy of the upper crust or the effects of pore pressure and compaction.

Numerical models of restraining bends

Numerical modelling to simulate restraining-bend deformation has been carried out by Rodgers (1979), Segall & Pollard (1980), Du & Aydin (1995) and Brankman & Aydin (2004). Rodgers (1979) approximated the Big Bend in the San Andreas fault zone by a series of straight dislocation surfaces, and calculated the displacement and stress fields for each fault segment. The vertical displacement fields resemble the general topographic features of the Transverse Ranges, suggesting that the range topography is bend-related. The north–south principal stress orientations also correspond with earthquake and other observations. Segall & Pollard (1980) and Du & Aydin (1995) simulated fracture patterns within restraining bends using different fault geometries.

Tectonic setting and description of natural bends on transform faults separating oceanic plates

Objectives

In this and following sections, I will describe the formation of both releasing and restraining bends in the five types of natural, plate-boundary-zone settings summarized in Figure 1 & Table 1. There are two reasons for including these maps and descriptions. First, regional maps are necessary to illustrate the presence of paired bends, which, as previously noted by Swanson (2005) are commonly not recognized as genetically linked pairs of releasing- and restraining-bend structures (Figs 10 & 11). The tendency in previous discussions of either releasing (Aydin & Nur 1982; Mann *et al.* 1983) or restraining bends (Segall & Pollard 1980; McClay & Bonora 2001) has been to focus on individual structures rather than considering a longer segment of the fault that may be controlling an array of fault bends. Second, regional maps are more valuable to illustrate the close relationship between fault patterns within the strike-slip system and large-scale plate motions, a relationship which is particularly well known from plate modelling studies and

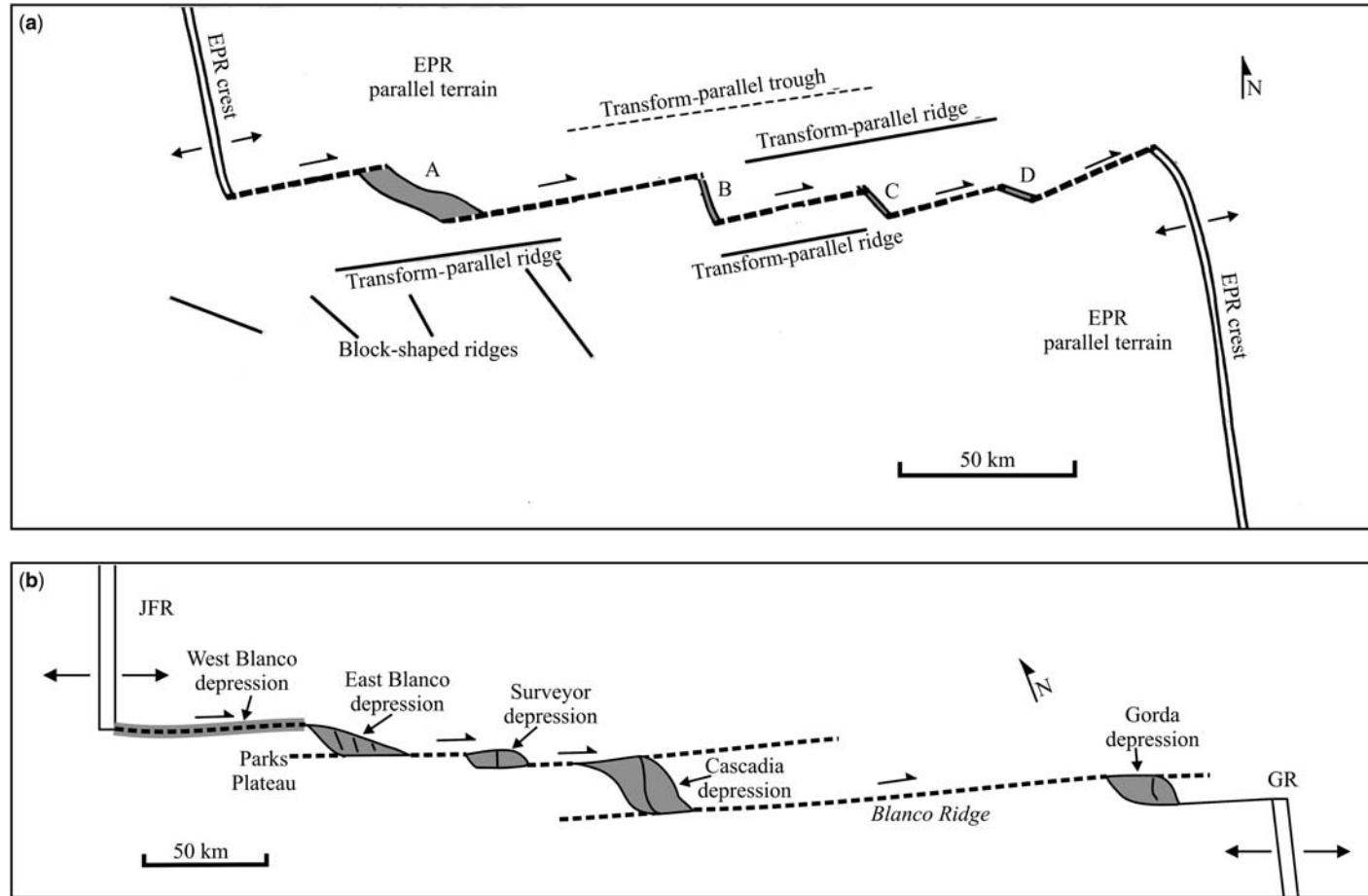


Fig. 14. (a) Array of four pull-apart basins found along the Siqueiros transform on the fast-spreading (63 mm/a) East Pacific Rise (EPR), modified from Fornari *et al.* (1989). (b) Array of four pull-apart basins found along the Blanco transform offsetting the fast-spreading (60 mm/a) Juan de Fuca (JFR) and Gorda (GR) ridges, modified from Embley *et al.* (1987).

GPS-based geodesy. As illustrated in Figures 8 and 9, the angles defined between the small circle about the pole of rotation and the strike of the fault are critical for determining the locations of individual or multiple arrays of pull-apart basins located along boundary transforms (Le Pichon & Francheteau 1978; Garfunkel 1981; Mann *et al.* 1983; Bilham & Williams 1985; Parkinson & Dooley 1996). Regional maps of strike-slip faults are also important for making the point that paired bends are not common for indent-linked strike-slip faults within interior areas of continental crust (Cunningham 2007) but do occur along trench-linked strike-slip faults.

Oceanic transforms

This class of faults connecting the offset ends of the mid-oceanic spreading ridges (Fig. 2) has been studied in detail using a variety of methods, including earthquake seismology, high-resolution bathymetry, studies of magnetic anomalies, and direct bottom sampling (Fox & Gallo 1984; Garfunkel 1986; Searle 1986; Fig. 14). Oceanic transforms typically form linear, steep-sided valleys, with the actively deforming zone marked by a single, narrow belt of deformation, or 'principal displacement zone' (Fox & Gallo 1984; Searle 1986). This fault morphology differs from wider, more arcuate and multi-branched patterns of faults seen in strike-slip faults affecting continental crust (Figs 4–6 & 10–11). The simpler and more focused patterns of deformation along oceanic transform faults probably reflect the thinner and more isotropic nature of oceanic crusts of varying ages that is lacking in the amount of inherited fabrics and cross-cutting features present in continental crust (Fox & Gallo 1984).

Another interesting contrast between oceanic transforms and continental strike-slip faults is the predominance of releasing bends, including those that have formed short spreading ridges (Garfunkel 1986; Searle 1986; Fornari *et al.* 1989). Only a single transpressional ridge (Fig. 1c) has been identified on the Clipperton transform fault in the Pacific Ocean (Pockalny 1997; Table 2). The lack of discrete restraining bends means that paired bends characteristic of most continental faults (Figs 10–11) have also not been identified on oceanic transform faults.

Pull-apart array on the Siqueiros transform, Pacific Ocean

Fornari *et al.* (1989) identified an array of four pull-aparts, or 'intra-transform spreading centres' on the Siqueiros transform (Fig. 14a). The pull-apart shapes are similar to those seen in continental settings: two of the basins exhibit lazy-Z morphology

between non-overlapping master faults; the other two basins are more orthogonal to the fault trace and rhomboidal in morphology. Geological data indicate that the basins formed from west (older basins) to east (younger basins) over the past 1 Ma. The stability of the East Pacific Rise for the past 1.5 Ma argues against a small change in plate motion to form the basins, as shown schematically in Figure 7a. Fornari *et al.* (1989) propose that the basins originated by the interaction of small melt anomalies in the mantle with lithospheric fractures within the transform fault.

Pull-apart array on the Blanco transform, Pacific Ocean

Detailed seismological, bathymetric and side-scan sonar mapping of the Blanco transform revealed an array of five, rhomboidal pull-apart basins arranged along the 350-km length of the transform that offsets the Gorda and Juan de Fuca spreading ridges in the NE Pacific Ocean near Washington State (Dziak *et al.* 1991; Embley & Wilson 1992; Fig. 14a). The geometry of the basins (Fig. 14b) is strikingly similar to rhomboidal pull-aparts formed in continental settings (Fig. 4). The centrally located and largest rhomboidal basin, the Cascadia depression, exhibits inwardly facing back-tilted fault blocks and is inferred to contain a short spreading centre active over the past 5 Ma. The basins have formed over the past 5 Ma as the result of the continual reorientation of the Blanco transform fault in response to changing plate motion. Plate reconstructions by Embley & Wilson (1992) to explain the formation of the pull-apart array favour a complex model involving propagating ridges instead of a simpler change in motion across a single, pre-existing, linear transform fault (cf. Figure 8a).

Transpressional ridge on the Clipperton transform, Pacific Ocean

Pockalny (1997) describes a currently active transpressional ridge using high-resolution bathymetry and earthquake data. He attributes the tectonic origin of transpression to a small change in the spreading direction along the East Pacific Rise, starting about 0.4–0.5 Ma.

Tectonic setting and description of natural bends on long, strike-slip faults separating continental plates

San Andreas fault system

Tectonic setting. The San Andreas fault system, extending 2100 km from the mouth of the Gulf of California to Cape Mendocino in northernmost

California, is undoubtedly the world's best-studied continental-plate-boundary strike-slip fault system (Woodcock 1986; Page 1990; Brown *et al.* 1991; Fig. 15 & Table 1). In this section, I summarize the current state of knowledge of restraining and releasing bends compiled from literature on the San Andreas fault system, with particular emphasis on their tectonic settings relative to the small circles

of rotation between the North America and Pacific plates (Fig. 15). I also review evidence for the existence of adjacent, 'paired bends' (Fig. 11) or regional 'sidewall ripouts' (Fig. 12) within the San Andreas fault system, whose occurrence and significance may be underestimated (Swanson 2005). For reviews of the tectonic origin of the San Andreas fault system, its offset history, and

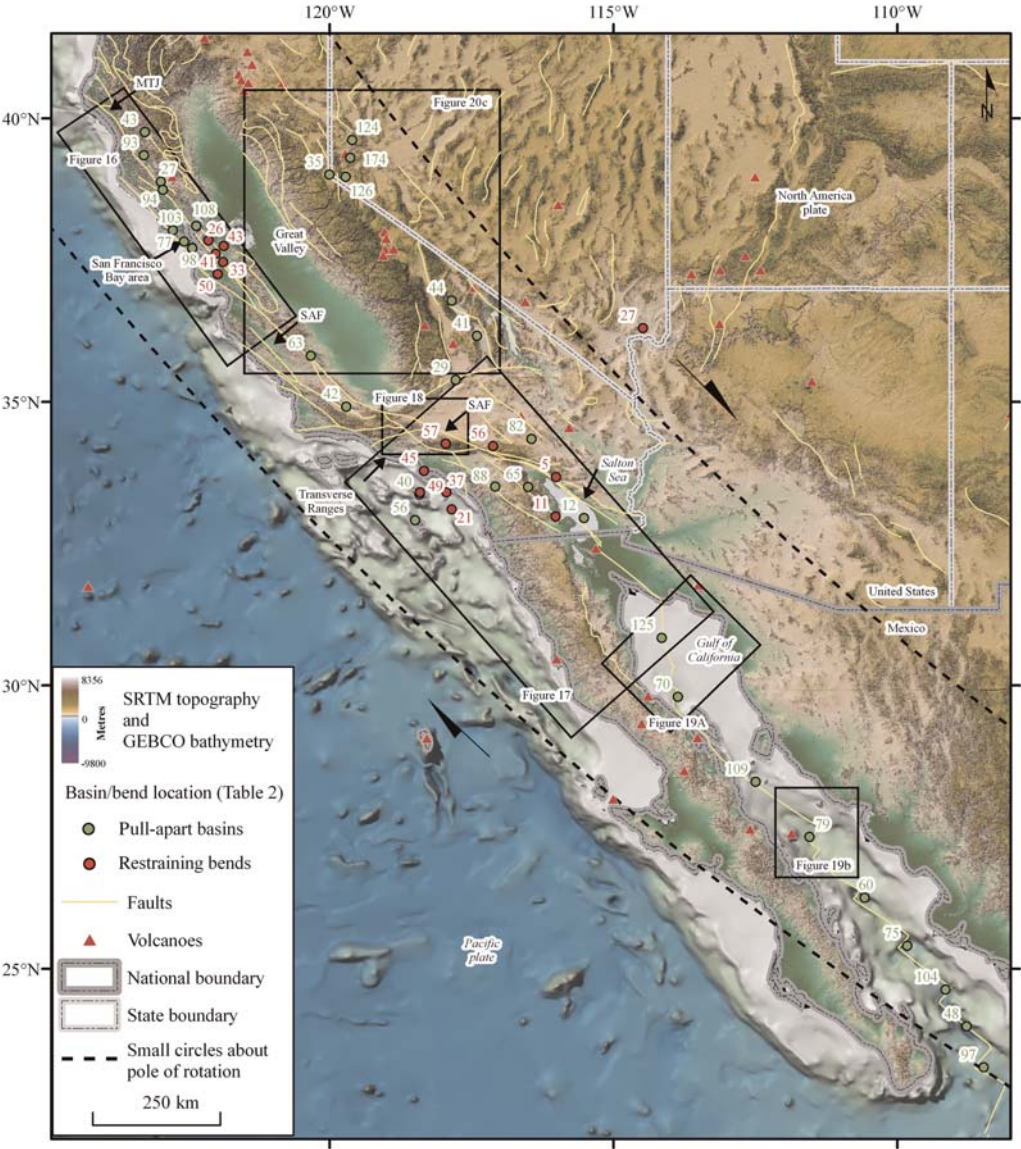


Fig. 15. Regional tectonic map of the San Andreas strike-slip fault system in California and the Gulf of California, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from the compilation by Levine (1995). Black lines represent small circles of rotation, about the pole of rotation, proposed by Argus & Gordon (2001). Boxes indicate areas of more detailed maps in Figures 16–20.

its palaeogeography, the reader is referred to Page (1990) and Powell & Weldon (1992).

The pole of North America–Pacific plate motion is distant and defines small circles roughly parallel to the concave-eastward traces of the San Andreas fault system (Argus & Gordon 2001; Fig. 15). The traces of the San Andreas fault system are smooth and anastomosing, and can be followed as continuous features for hundreds of kilometres. Analysis of GPS and plate kinematic data indicates that the overall character of the San Andreas Fault is slightly transpressional in character, with the fault strands accommodating about 3 mm/a of orthogonal shortening and about 37 mm/a of right-lateral strike-slip faulting (Wesnousky 2005). This slight transpressional character of the fault can account for fault-parallel faults and folds that parallel straight traces of the fault in central and northern California that are not characterized by restraining bends (Mount & Suppe 1987). Most of the 300–450 km of total right-lateral strike-slip displacement has occurred on the main San Andreas fault zone; parallel zones of strike-slip faults have offsets ranging from a few kilometres to 40 km (Powell & Weldon 1992; Wesnousky 2005).

The San Andreas fault zone can be more precisely described as the boundary separating the Pacific plate and the Sierran microplate, the latter of which includes the Great Valley and Sierra Nevada of eastern California and moves about 12 mm/a relative to the North American plate. This motion is occurring mainly in a zone of trans-tensional, right-lateral deformation separating the eastern edge of the Sierra microplate from the North American plate in the Great Basin (Unruh *et al.* 2003; Wesnousky 2005).

Bends on the San Andreas Fault. Three tectonic settings for the formation of San Andreas fault bends compiled on Table 2 can be defined:

1. a region in northern California that extends from north of the Transverse Ranges to Cape Mendocino where the traces of the strike-slip system are close to parallel or slightly convergent ($<5^\circ$) with respect to the small circles of rotation;
2. a region of southern California where the traces of the strike-slip fault system are more convergent ($5\text{--}15^\circ$) with respect to the small circles of rotation; this region contains major restraining bends at the topographically elevated Transverse Ranges and San Bernardino Mountains; and
3. a region of southernmost California, northern Mexico and the Gulf of California where the traces of the strike-slip fault system are divergent ($5\text{--}10^\circ$) with respect to the small circles of rotation; this region of the broad, low-lying

Imperial Valley of California and northern Mexico and the Gulf of California is characterized by *en échelon* pull-apart basins that lack associated restraining bends (Fig. 15).

Strike-slip bends of the slip-parallel region of northern California

Regional fault pattern of bends. The dominant bend structure in northern California is the lazy-S-shaped, Santa Cruz Mountains gentle restraining bend defined by a left-step (fault separation of 20 km) in the San Andreas fault zone south of San Francisco Bay (Fig. 16). Burgmann *et al.* (2006) have used GPS and the interferometric synthetic aperture radar (INSAR) method in the San Francisco Bay area to isolate vertical tectonic uplift rates at sub-mm per year precision. This survey indicates tectonic uplift at about 1 mm/a over a broad area of the Santa Cruz restraining bend (Fig. 16).

North of the bend, the San Andreas Fault follows the coastline and is marked by a prominent but narrow (c. 2–3 km) fault valley that is occupied by several isolated releasing rhomboidal bends at the Golden Gate and in Olema Bay with fault separations less than 2 km and poorly known transverse normal faults (Zoback *et al.* 1999; Wakabayashi *et al.* 2004, 2007). Parsons *et al.* (2005) propose a more complex fault-wedge basin model (Biddle & Christie-Blick 1985) for the simple pull-apart model shown by Zoback *et al.* (1999) and Wakabayashi *et al.* (2004, 2007).

This strand of the San Andreas Fault ruptured for several metres over a minimum onshore distance of 220 km in 1906, from the central part of the gentle Santa Cruz bend to the offshore region of the fault in northern California, to produce an *M*7 earthquake. This strand accommodates 40 mm/a of Pacific–North America motion, or about 90% of the total motion between the Pacific plate and the Sierran/North America plate (Argus & Gordon 2001).

As the fault passes seaward at Point Arena and extends northward to the most recently developed part of the fault near Punta Gorda and beyond to the Mendocino triple junction, a significant deviation occurs in the trace of the San Andreas Fault as the fault bends to the NE and defines a fault stepover distance of 50 km. This part of the fault is characterized by an unusually large component of vertical slip (0.5 cm/a, NE side up) (Brown 1991). Merritts & Bull (1989) have attributed this enhanced vertical slip to the recent passage of the Mendocino triple junction, but an alternative explanation is that part of the fault forms a gentle releasing bend that is largely located on fault splays offshore (McCulloch 1987) and defines the

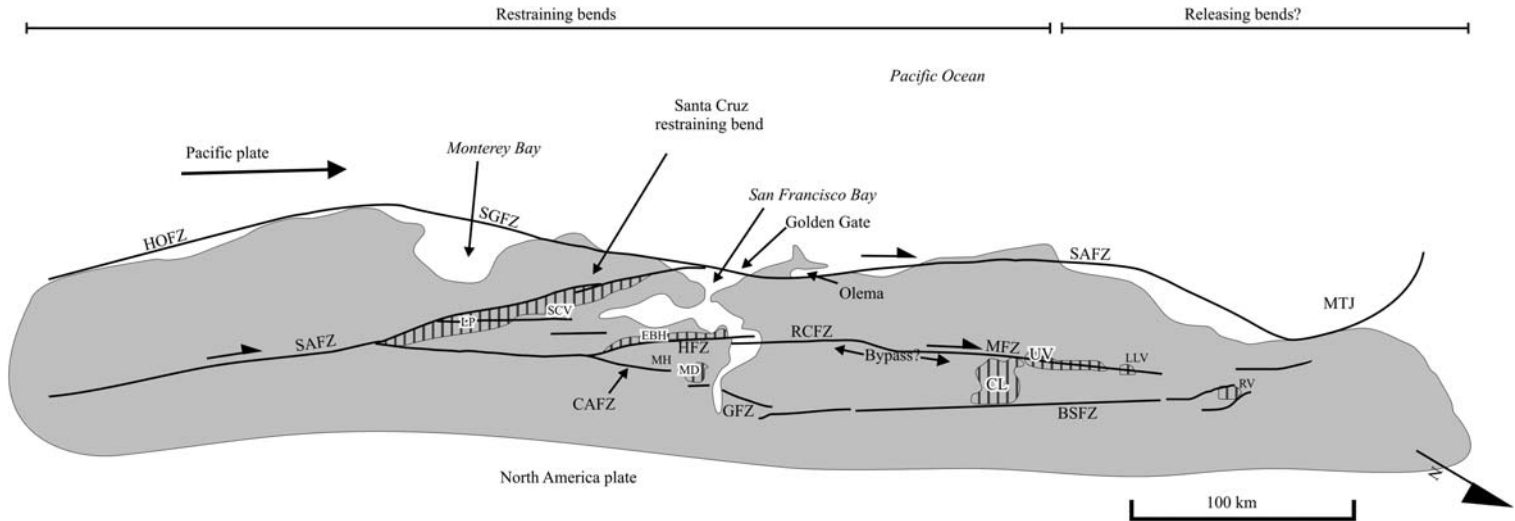


Fig. 16. Schematic map of the active traces of the San Andreas fault system in northern California, modified from Wesnousky (2005), McLaughlin & Nilsen (1982), Burgmann *et al.* (2006) and Aydin & Page (1984), and their relation to North America–Sierra Nevada–Great Valley microplate–Pacific plate motions, from Argus & Gordon (2001). See the regional tectonic map in Figure 15 for the location. An inferred paired of restraining and releasing bends is indicated on the San Andreas fault zone, along with a possible bypass fault system across the base of the paired bend (Calaveras–Hayward–Rogers Creek–Maacama fault zones). Note how the coastline of northern California mirrors the coastal and offshore paired-bend structure. HOFZ, Hosgri fault zone; SGFZ, San Gregorio fault zone; SAFZ, San Andreas fault zone; HFZ, Hayward fault zone; CAFZ, Calaveras fault zone; GFZ, Green Valley fault zone; RCFZ, Rodgers Creek fault zone; BSFZ, Bartlett Springs fault zone; LP, Loma Prieta; SCV, Santa Clara Valley; EBH, East Bay Hills; MH, Mission Hills; MD, Mount Diablo; CL, Clear Lake; MFZ, Maacama fault zone; UV, Ukiah Valley; LLV, Little Lake Valley; RV, Round Valley; MTJ, Mendocino triple junction.

coastline where Holocene uplift rates of 0.5 cm/a have been measured (Brown 1991; Fig. 16). I propose that this bend may represent the releasing-bend (cf. Fig. 1d) component of the Santa Cruz restraining bend to the south, since both share approximately the same amount of fault separation (Fig. 16).

East of the San Andreas Fault is a slightly segmented series of faults that diverges from the San Andreas at the Calaveras fault zone near the SW end of the Santa Cruz gentle restraining bend (Fig. 16). This second strand of the fault system, which is less continuous than the San Andreas strand 75 km to the west includes the Calaveras, Hayward, Rodgers Creek and Maacama faults (McLaughlin & Nilsen 1982; Wakabayashi 2007). In northern California, the northern strands of both this fault and the parallel San Andreas Fault represent the youngest part of the system that has propagated northward in the wake of the Mendocino triple junction (Henstock *et al.* 1997; Wakabayashi *et al.* 2004, 2007; Fig. 16). A third parallel strand representing the easternmost active strands of the San Andreas fault zone is even more discontinuous, and consists of the Green Valley and Bartlett Springs fault zone (McLaughlin & Nilsen 1982; Fig. 16).

Argus & Gordon (2001) note that the Calaveras–Maacama trend and the parallel San Andreas trend are very close to parallel to the small circle of motion between the Pacific and Sierran microplates (Fig. 15). An interesting regional structural correlation is that restraining bends are mainly confined to the southern part of the strike-slip system in a 120-km-long zone adjacent to the large Santa Cruz gentle restraining bend (Fig. 16). These much smaller bends include the Mission Hills gentle restraining bend connecting the Hayward and Calaveras faults (Manaker *et al.* 2005); the Mount Diablo gentle restraining bend between the Concord–Greenville fault zones (Wakabayashi 2007); and the East Bay Hills gentle restraining bend along strands of the Hayward fault zone (Aydin & Page 1984). Burgmann *et al.* (2006) has identified recent topographic uplifts with all of these bends using the INSAR method.

The northern part of this strike-slip system from San Pablo Bay (the easternmost extension of San Francisco Bay) northward to Mendocino appears to be transtensional and characterized by a lack of restraining bends and the presence of either pull-apart basins or fault wedge basins (McLaughlin & Nilsen 1982; Fig. 16). This inland region contains small pull-aparts or fault wedge basins ranging in fault separation from 1.5 to 2 km, and are either occupied by lakes or shallow-marine areas or have been mapped as outcrop areas of Late Neogene continental sedimentation

(McLaughlin & Nilsen 1982; Wakabayashi *et al.* 2004, 2007).

These basins occurring at *en échelon* stepovers or aligned along the fault traces or at complex fault traces include:

1. the rhomboidal San Pablo pull-apart underlying San Pablo Bay (Parsons *et al.* 2003);
2. the Little Sulfur Creek, Ukiah Valley, and Little Lake Valley basins, pull-aparts or fault wedge basins aligned along the Maacama fault zone; and
3. the Clear Lake, Potter Valley, and Round Valley basins along the Lake Mountain, Bartlett Springs and Round Valley fault zones (McLaughlin & Nilsen 1982; Fig. 16).

Consistent with its transtensional setting, the Clear Lake Basin is also characterized by volcanic and geothermal activity, beginning at about 0.6 Ma, and subsidence rates of 1.7 mm/a for the past 0.6 Ma (Hearn *et al.* 1988).

Paired-bend interpretation. A simple interpretation of the regional fault pattern in Figure 16 is that there are paired bends on the San Andreas fault strand to the west, the Calaveras–Hayward–Rodgers Creek–Maacama fault zone in the centre and the Concord–Greenville–Lake Mountain–Bartlett Springs and Round Valley fault zones to the east. All three faults exhibit a convex westward shape that is most pronounced in the case of the San Andreas fault and is a result of the Santa Cruz gentle restraining bend and perhaps its paired bend located off the coast of northern California. Given the direction of interplate slip, the northern part of this arcuate belt of three parallel strike-slip faults north of San Pablo Bay is characterized by pull-apart or fault wedge basins along their traces, along with volcanic and geothermal activity centred on those basins.

The largest estimate of right-lateral displacement on the eastern strands of the fault system is 24 km on the Calaveras fault zone in contrast to the San Andreas strand that has a right-lateral offset on basement terranes of 450 km (Wesnousky 2005). The low offsets on the eastern strands are consistent with the idea that these faults initiated fairly recently – about 8 Ma according to McLaughlin & Nilsen (1982) – and have propagated northward in the wake of the Mendocino triple junction since that time (Henstock *et al.* 1997). Geological and fission-track studies of the Santa Cruz Mountains bend indicate that its earliest uplift occurred about 12 Ma (Middle Miocene), and since that time about 2 km of exhumation has occurred at the bend (Anderson 1990; Burgmann *et al.* 1994).

In the framework of paired bends, the eastern fault strands may represent a 'bypass fault' that is reducing the paired-bend asperity, or 'sidewall ripout' (Swanson 2005, Fig. 12b), developed along the San Andreas fault zone over at least the past 12 Ma and given the large offset, perhaps much longer. Because the inferred eastern bypass fault system is straighter and parallel along its length to the small circles of plate rotation, its activation would increase the efficiency of the strike-slip system as described by Wesnousky (1988). It is puzzling why the two parallel strands of the San Andreas fault system inherit the same westward convexity that in turn localizes aligned zones of restraining bends to the south of San Francisco Bay on the San Andreas fault zone and to the east of the bay on the Hayward fault zone (Fig. 16). Two possibilities are that these trends are controlled by inherited basement structures (Fig. 12c), or that they reflect the sidewall ripout process proposed by Swanson (2005; Fig. 12b). It seems unlikely and coincidental that two parallel strike-slip fault strands would align themselves by linkage of P- and R-shears as shown in Figure 12a.

Strike-slip bends of the convergent region of southern California

Regional fault pattern of bends along the Big Bend of the San Andreas fault zone and northern Gulf of California. The dominant bend structure in southern California is the lazy-S-shaped, Transverse Ranges gentle restraining bend, defined by a left-step (fault separation of 20 km) in the San Andreas fault zone north and east of Los Angeles (Fig. 15). In this region, the small circles of plate rotation are oblique by 15–20° to the fault traces and thereby provide a tectonic setting for the formation of the largest restraining-bend complex and high topography (3502 m in the San Bernardino Mountains) in California (Argus & Gordon 2001; Fig. 17a). As the San Andreas fault trace approaches the Salton Sea, it curves, straightens, and assumes a more slip-parallel orientation, and it is characterized by structures that are either slip-parallel or slightly transpressional (Bilham & Williams 1985; Fig. 17a). Because of excellent outcrops in a dry desert climate, this segment of the fault is the setting for the classic outcrop-based mapping study of strike-slip transpressional deformation by Sylvester & Smith (1976) in the Mecca Hills north of the Salton Sea (Fig. 17b).

These areas of localized fault transpressional are best characterized as gentle bends connected by zones of pure slip faulting (Bilham & Williams 1985; Fig. 17b). The kinematic and morphological properties of each segment shown in Figure 17b

depend entirely on the fault strike, despite very subtle changes in fault strike that are as small as 3°. Note that these fault strike changes are so subtle that the distinction between a gentle fault bend and a linear transpressional uplift (Fig. 1c) may not be clear.

At a point west of the Salton Sea, the San Andreas Fault abruptly curves to the SE and assumes a SE trend oblique in the extensional sense to the small circles of plate rotation (Fig. 17a). Along this segment of the fault that extends from the Salton Sea to the mouth of the Gulf of California, the strike-slip system is transtensional and therefore provides the tectonic setting for the formation of an *en échelon* array of 11 active pull-apart basins exhibiting the complete range of developmental morphologies that are summarized on Figures 1b & 7 (Mann *et al.* 1983).

Pull-apart basin morphologies include: spindle-shaped types (Upper Wagner Basin on Fig. 18a; Henyey & Bischoff 1973); rhomboidal types (Delfin Basin in Fig. 18a; Henyey & Bischoff 1973; Persaud *et al.* 2003) and rhomboidal type with oceanic spreading centres (Guaymas Basin on Figure 18b; Bischoff & Henyey 1974). The spindle-shaped basins have small fault overlaps and incipient coalescence with adjacent basins, reflecting their recent nucleation in the sub-sea-level areas SE of the Salton Sea area near the top of the convex eastward arc in the San Andreas Fault (Fig. 17). Larger, more isolated rhomboidal basins indicative of older basins formed by larger strike-slip offsets are found to the SE in the submarine Gulf of California (Fig. 15).

Transtensional structures and sub-sea-level topography within the Imperial Valley and Gulf of California reflect both the presence of pull-apart basins along the trend of the San Andreas Fault and also the activity of large normal faults bounding the walls of the valley. Umhoefer & Stone (1996) have pointed out that some of the transtensional motion is accommodated or 'partitioned' on the valley-parallel faults which form large escarpments bounding both the Imperial Valley and Gulf of California, similar to that discussed by Garfunkel (1981) in the Dead Sea, which is shown schematically in Figure 9c.

Regional pattern of bends in the California borderlands. The pattern of bends in the offshore California borderlands mirrors the pattern of bends along the onland San Andreas fault zone (Legg *et al.* 2007; Fig. 17). Most restraining bends are concentrated in the NW area of the California borderlands near the city of Los Angeles and the Transverse Ranges, whilst most releasing bends and straight strike-slip fault segments are concentrated in the SE areas near the city of San Diego

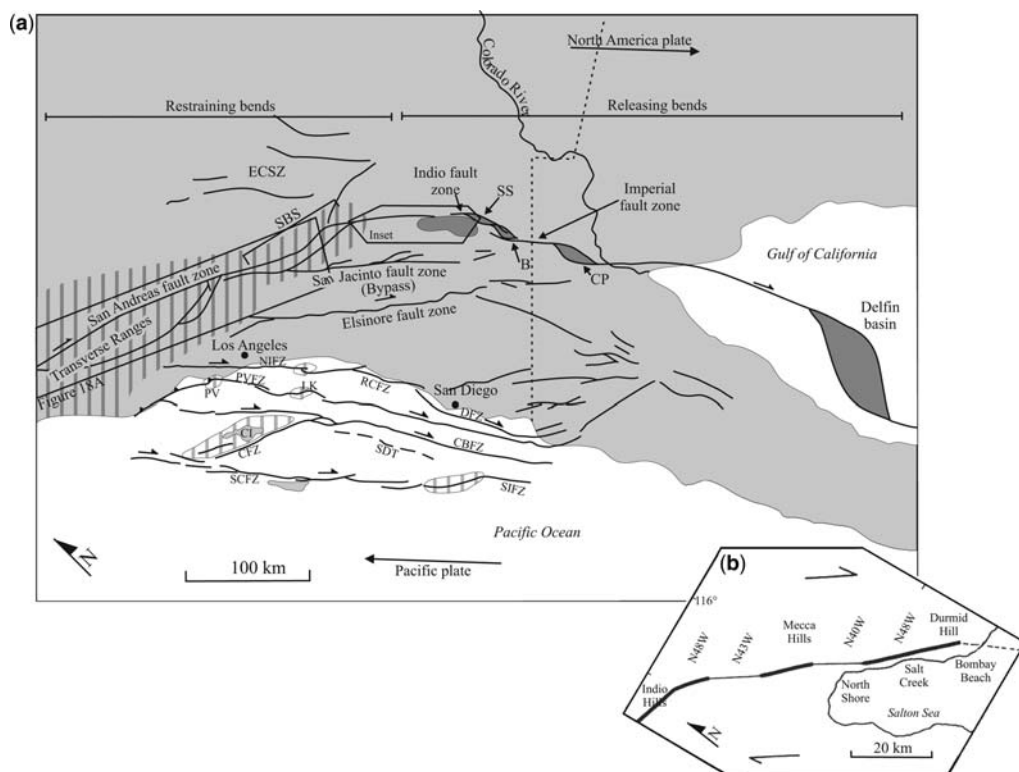


Fig. 17. (a) Schematic map of the active traces of the San Andreas fault system in southern California, from Wesnousky (2005), Persaud *et al.* (2003) and Dixon *et al.* (2002) (see the regional map in Fig. 15 for location). Schematic map of the active traces of the San Andreas fault system in the California borderlands and coastal southern California, modified from Legg *et al.* (2004), and their relation to plate motions, from Argus & Gordon (2001). Inferred paired restraining- and releasing bends are indicated for the three fault systems in the NE part of the map. Note how the coastline of southern California mirrors the paired-bend structure of the coastal and offshore fault systems. The San Clemente–San Isidro fault system in the SW part of the map and furthest offshore does not display the paired-bend pattern. (b) Map of the trace of the San Andreas fault zone to the NW of the Salton Sea, showing the close relationship between the strike of the fault and localized restraining bends (heavy lines) characterized by topographic uplift (modified from Bilham & Williams 1985).

and the Mexican border (Fig. 17). The strike-slip faults define a broad arc with restraining bends in the NW area, including the Palos Verdes, Lasuen Knolls and Santa Catalina Island gentle restraining bends (Ward 1994; Mann & Gordon 1996; Fisher *et al.* 2004; Legg *et al.* 2007). Pull-aparts are relatively rare and small in size in the California borderlands relative to restraining bends (Legg *et al.* 2007).

Paired-bend interpretation. The arcuate and sub-parallel strike-slip faults in southern California exhibit a convex eastward shape that is most pronounced for the Transverse Ranges restraining bend and its paired releasing bend, starting in the Salton trough area and extending into the northern Gulf of California (Fig. 17). The convex NE

curvature of faults in the California borderlands is less than that along the San Andreas fault zone, but still results in the Santa Catalina Island restraining bend parallel to the Transverse Ranges bend (Fig. 17). Southward-curving strike-slip fault trends produce transensional pull-apart that are most pronounced along the San Andreas fault zone in the Salton Trough and northern Gulf of Mexico. As in the case of the bends of northern California, it is puzzling why parallel sets of strike-slip faults along the San Andreas and California Borderlands trends exhibit the same sense of curvature across such a wide area of the plate-boundary zone. Two possibilities are that these trends are reactivating inherited structures in the basement (Fig. 12c), or alternatively that they reflect the sidewall ripout process proposed by Swanson (2005; Fig. 12b).

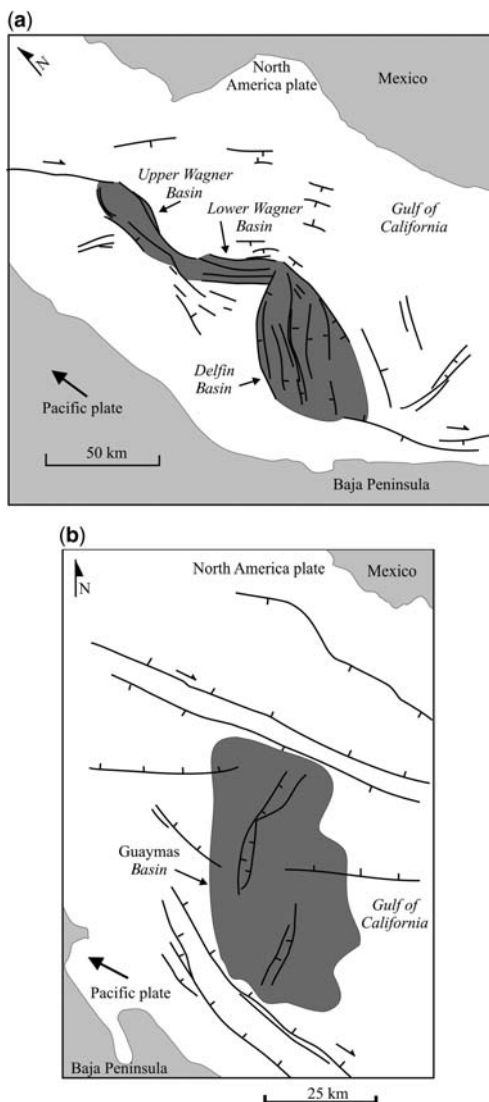


Fig. 18. (a) Schematic map of the Upper Wagner, Lower Wagner and Delfin pull-apart basins in the northern Gulf of California, modified from Henyey & Bischoff (1973) (see the regional map in Figure 15 for the location). The direction of plate motion is from Argus & Gordon (2001). (b) Schematic map of the Guaymas pull-apart basin in the central Gulf of California, modified from Bischoff & Henyey (1974).

Development of the Transverse Ranges restraining bend. Powell & Weldon (1992) proposed that the earliest Transverse Ranges bend was active from the Early to Mid-Miocene (18–13 Ma), and that this early bend accumulated up to 110 km of right-lateral displacement. This early phase of bend formation led to the cessation of slip on the San

Andreas Fault, and transfer of slip to the San Gabriel fault zone from the Early Miocene (14–12 Ma) to Early Pliocene (4–5 Ma) (Crowell & Link 1982; Fig. 19a). Seismic profiling by May *et al.* (1993) has shown that the San Gabriel Fault is a low-angle normal fault (Fig. 19b) upon which 13.5 km of Mio-Pliocene marine and non-marine sedimentary fill has been deposited (Fig. 19c). As the San Gabriel Fault shifts to a more SE strike, the basin disappears and the fault localizes the uplift in the San Gabriel Mountains (Fig. 19a).

Swanson (2005) has interpreted this fault pattern and kinematic history as a sidewall ripout: an Early Miocene shift from the San Andreas to the San Gabriel Fault defined the ripout with its characteristic plan-view fault pattern. When the ripout was active from the Early Miocene to Early Pliocene, a basin formed behind the ripout as a releasing bend at the trailing edge of the block, and the San Gabriel mountain range formed at the leading edge. As motion shifted back to the San Andreas Fault, the basin and restraining bend were abandoned to allow a straighter fault trace (Fig. 19a).

The latest phase of Transverse Ranges uplift occurred as motion was re-established on the San Andreas fault trace in Late Pliocene to Recent times, and is recorded by Late Pliocene and Quaternary folds, thrust faults and reverse faults (Namson & Davis 1988). Although an earlier late Mid-Eocene to Late Oligocene folding and uplift event affected the Transverse Ranges, the latest event is responsible for most of the present structural relief and topography in the area. In a study of fission-track ages across the central Transverse Ranges, Blythe *et al.* (2000) found that the most recent phase of uplift in the San Gabriel Mountains thought to be linked to its restraining-bend history started at about 7 Ma, with an accelerated phase beginning at 3 Ma. The topographically higher San Bernardino Mountains to the east have experienced less dissection than the San Gabriel mountains, suggesting that the most recent phase of cooling and bedrock uplift began about 3 Ma ago. Spotila *et al.* (2001) used apatite ages to constrain the data indicating that a small fault sliver in the San Bernardino restraining bend was exhumed 3–6 km at rates of 5 mm/a over the past 1.8 Ma.

Rhomboidal pull-aparts like the Wagner, Delfin, and Guaymas basins in the northern and central Gulf of California formed at about 3 Ma ago (Henyey & Bischoff 1973; Bischoff & Henyey 1974; Boehm 1984) and are therefore roughly contemporaneous with the main phase of uplift and shortening on the Transverse Ranges gentle restraining-bend uplift (Fig. 17). The younging trend in pull-apart basin opening from the northern Gulf of California to the Salton Trough is also supported by the morphology of the spindle-shaped basins beneath the

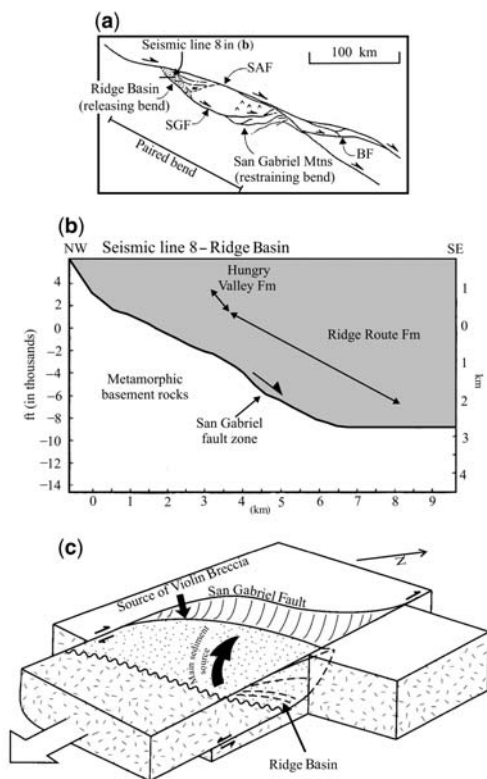


Fig. 19. (a) Schematic map of the active traces of the San Andreas fault system in southern California, from Swanson (2005) (see the regional map in Figure 15 for the location). The location of the onland, multi-channel seismic line crossing the Ridge Basin and the bounding San Gabriel fault low-angle normal fault zone, is shown in (b). Prior to this geophysical information, the San Gabriel fault zone was interpreted as a high-angle strike-slip fault zone (Crowell 1974a, b; Crowell & Link 1982). (b) Half-graben structure formed by mainly low-angle, normal slip on the San Gabriel Fault during Late Miocene times (modified from May *et al.* 1993). Wedge-shaped sedimentary deposits of the Ridge Basin above the San Gabriel normal fault are 5 km thick and consist of alluvial-fan, lacustrine and shallow marine sedimentary rocks. (c) Block diagram illustrating the low-angle normal fault formed at a releasing bend along the Mio-Pliocene San Gabriel fault zone, from May *et al.* (1993). This releasing bend was inferred by Swanson (2005) to have formed a trailing extensional ramp system or releasing bend (i.e. Ridge Basin) adjacent to the San Gabriel leading thrust system or restraining bend (cf. map in (1d)). In the terminology of this paper, the adjacent bends would be called a paired bend, as shown in Figure 14a. The modern San Andreas fault zone bypassed this ancient paired bend about 4–5 Ma ago, and subsequent transpressional deformation has exhumed the Ridge Basin in its present-day outcrop area.

Salton Trough, which started to form less than 1 Ma ago (Fuis *et al.* 1982) suggested started to form less than 1 Ma ago. Therefore, the youngest area of restraining-bend deformation in the San Bernardino Mountains uplifted from 1.8 Ma ago (Spotila *et al.* 2001) is paired or adjacent to the youngest area of spindle-shaped pull-apart basin formation in the Salton Trough, active from about 1 Ma ago. Older areas of restraining-bend formation in the San Gabriel Mountains (Blythe *et al.* 2000) are paired with older areas of rhomboidal pull-apart opening in the Gulf of California (Persaud *et al.* 2003; Fig. 17).

One structural explanation to explain the symmetry and synchronous development of the paired bend is that the excess volume of deformed crust 'piling up' at the Transverse Ranges must be balanced by a volume deficit by thinning the crust in the adjacent, along-strike area of the Salton Sea pull-aparts (Fig. 17). Since mass balance of crust is preserved, the area of thinned crust is equivalent to the thickening in the restraining bend, and therefore the approximate shapes and master-fault separations would remain similar.

Strike-slip fault bends in the Walker Lane transtensional zone

A 100-km-wide diffuse transtensional deformation along Walker Lane fault zone and Eastern California shear zone occurs at the eastern edge of the Sierra microplate, and has accommodated a maximum of 150 km of right-lateral displacement since its inception (Oldow 2003; Wesnousky 2005; Fig. 20a). Unruh *et al.* (2003) and Wesnousky (2005) have both noted that the fault trends obliquely with respect to the small circles of rotation between the Sierran microplate and North America, and that this misalignment has produced a highly transtensional fault that accommodates about 11 mm/a of right-lateral shear (Fig. 20a).

Because of this misalignment, Unruh *et al.* (2003) view the 500-km-long central segment of the Walker Lane fault zone to be a releasing step-over composed of many individual rift basins and transtensional faults (Fig. 20b). Discrete pull-aparts within this larger pull-apart may include several small and steep-sided basins, including the classic Z-shaped Death Valley pull-apart basin (Burchfiel & Stewart 1966; Serpa *et al.* 1988; Cowan *et al.* 2003); Lake Tahoe Basin (Unruh *et al.* 2003); the Coso Basin and geothermal field (Lees 2002); and the Walker Lake pull-apart basin (Link *et al.* 1985). Wesnousky (2005) notes the presence of a 100-km-wide, rhomboidal stepover in the central Walker Lane that is deforming by rotation of three rectangular blocks bounded by antithetic

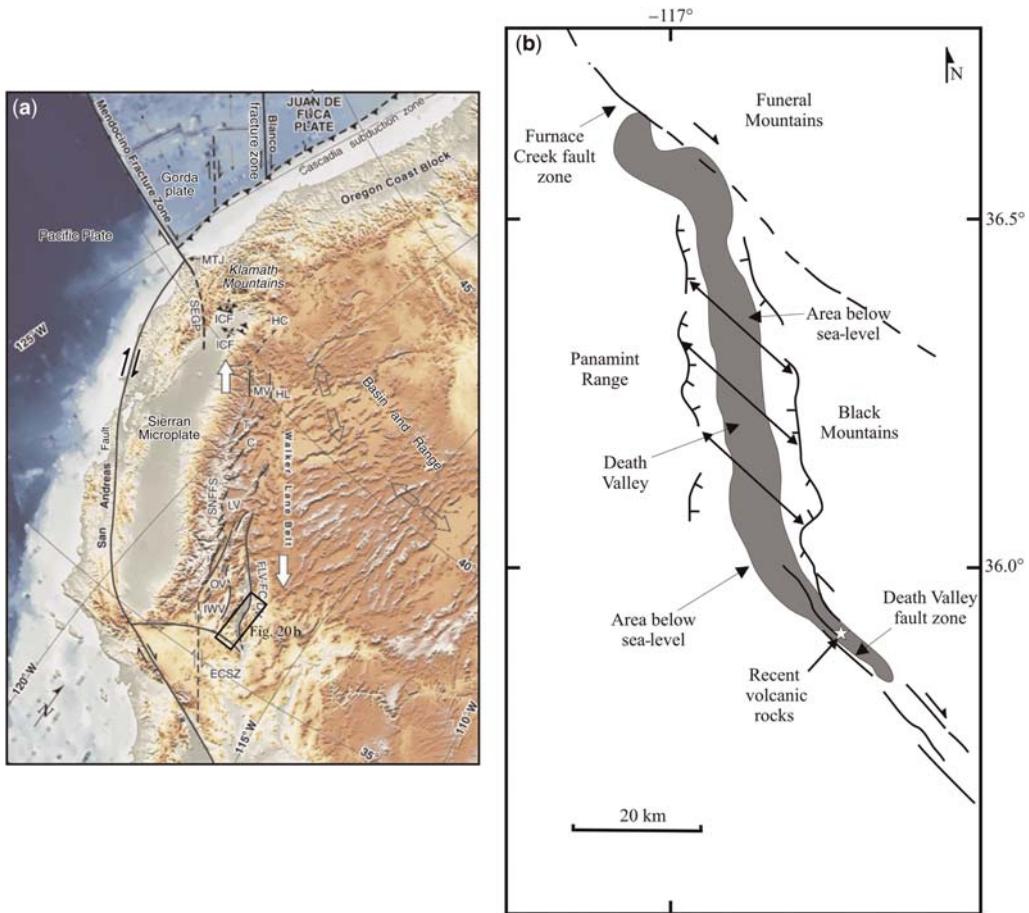


Fig. 20. (a) Oblique Mercator projection of the western Cordillera about the preferred Sierra Nevada–North America pole of rotation of Argus & Gordon (2001) (from Unruh *et al.* 2003). Direction of instantaneous Sierra Nevada–North America motion is vertical everywhere in the projection. Faults of the Walker Lane shear zone are transtensional. (b) Schematic map of the active, S-shaped Death Valley pull-apart basin along the Furnace Creek and Death Valley fault zones in southern California, modified from Christie-Blick & Biddle (1985). The grey area indicates part of the basin, from sea-level to a minimum elevation of 86 m below sea-level.

(left-lateral) strike-slip faults. He notes that the overall diffuse and transtensional character of Walker Lane may be analogous to an earlier stage in the structural development of the San Andreas fault system before sufficient slip accumulated to yield the now more concentrated San Andreas strike-slip fault zone.

Tectonic setting and geological development of bends along the Dead Sea strike-slip fault system

Tectonic setting. The Dead Sea fault system extends from its intersection with the East Anatolian strike-slip fault of SE Turkey in the north to

the Gulf of Elat and the Red Sea oceanic spreading centre in the south (Fig. 21). Like the San Andreas Fault, the Dead Sea Fault ranks as one of the world's best-studied strike-slip faults (Quennell 1956, 1958; Freund *et al.* 1970; Daeron *et al.* 2005; Gomez *et al.* 2007). In this section, I will focus on the current state of knowledge of its major restraining and releasing bends, with particular emphasis on their tectonic setting relative to the small circles of rotation between the Africa and Arabia plates. More detailed descriptions of the Dead Sea fault, including its origin, offset history and palaeogeography, are given by Hempton (1987), Garfunkel *et al.* (1981) and Gomez (2007).

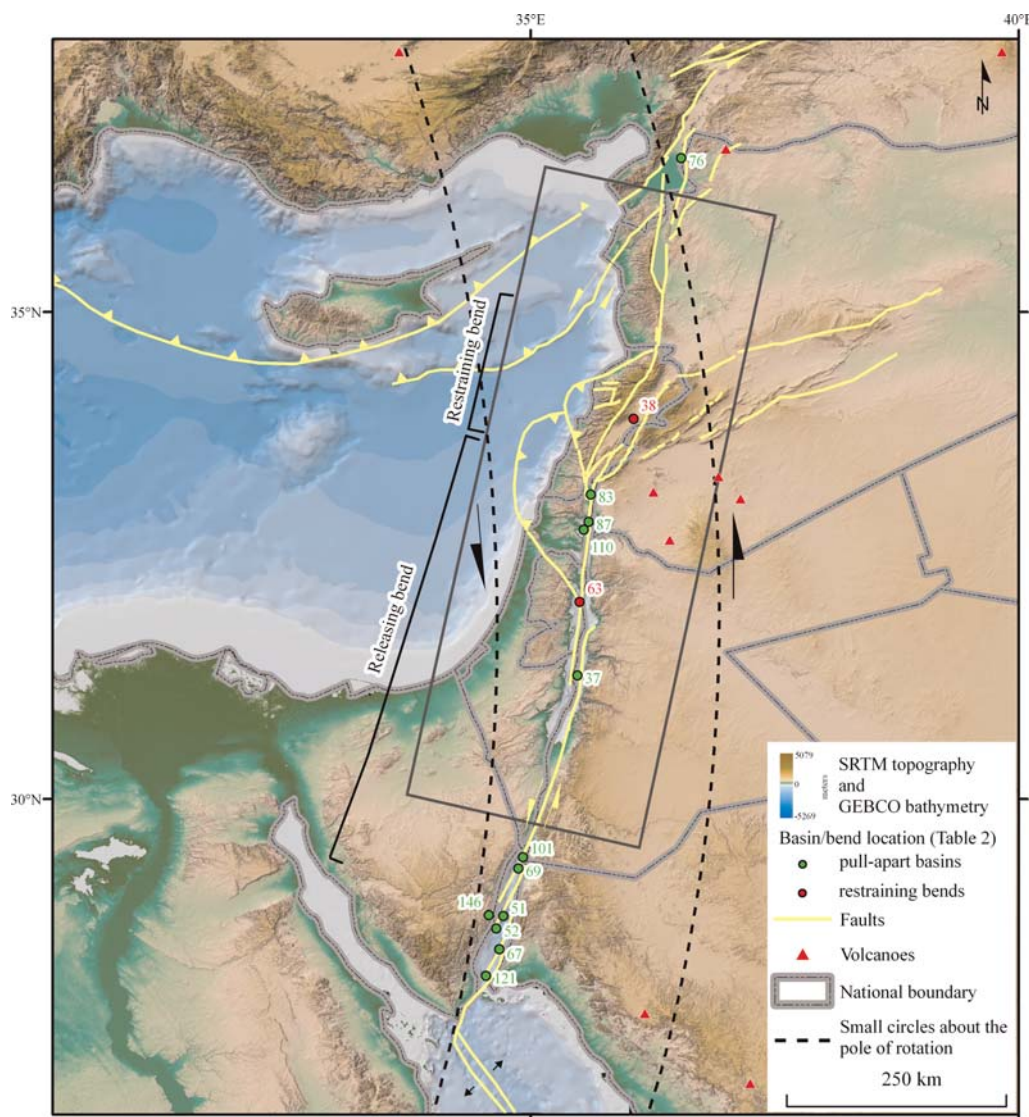


Fig. 21. Regional tectonic map of the Dead Sea strike-slip fault system in the Middle East, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults, modified from Gomez *et al.* (2006, 2007). Black lines represent small circles of rotation about the pole of rotation proposed by McClusky *et al.* (2003) using GPS observations. An inferred paired restraining- and releasing bend is indicated.

Unlike the multi-branched San Andreas and Walker Lane fault systems, the Dead Sea Fault is remarkably linear and focused on a single major strand along much of its length. The total amount of left-lateral offset (110 km) is known from geological studies (Freund *et al.* 1970) and aeromagnetic mapping (Hatcher *et al.* 1981) because the fault cuts at a high angle across Precambrian and younger

rocks. Geological studies indicate that offset occurred in two main phases. The first phase of *c.* 60 km of left-lateral displacement occurred during the Mid- and Late Miocene. The second phase began in the Early Pliocene and continues to the present-day. It may have been activated by an episode of 'ridge push' related to the onset of rifting and seafloor spreading in the Red Sea (Hempton 1987).

GPS studies indicate left-lateral strike-slip movement at rates of *c.* 5–8 mm/a with an increasing convergent component along the northern part of the fault (McClusky *et al.* 2003). Rates of motion also increase from 5.6 mm/a in the south to 7.5 mm/a along the northern segment of the fault (Fig. 21). These predicted, GPS-based small circles of rotation differ significantly from more fault-parallel small circles, based on a pole derived from plate modelling by Le Pichon and Francheteau (1978) and used by Garfunkel (1981) and Mann *et al.* (1983) in predicting zones of releasing and restraining bends along the fault.

Bends on the Dead Sea Fault. Gomez *et al.* (2006) has discussed the geological implications of the more precise GPS-based plate motion model on the observed bend structures along the fault. The Dead Sea fault system can be divided into three tectonic settings for restraining and releasing bends based on small circles:

1. a *c.* 400-km-long section of *en échelon* pull-apart basins from the Gulf of Elat through the Dead Sea pull-apart, Jordan River Valley and Sea of Galilee pull-apart; this section is roughly parallel to small circles in the Gulf of Elat, but becomes oblique and convergent by angles of 10–15° in the northern part of the segment;
2. a *c.* 200-km-long NE-striking, lazy-Z-shaped gentle restraining bend through the Mount Lebanon and Anti-Lebanon Ranges; this section is characterized by the maximum obliquity (*c.* 15–20°) of the small circles; and
3. a *c.* 250-km-long, north–south striking section in NW Syria and SE Turkey that contains one large, rhomboidal pull-apart basin (Brew *et al.* 2001).

Gomez *et al.* (2005) suggests that partitioning can explain the coexistence of pull-apart basins like the Ghab Basin within a predicted tectonic environment of tectonic transpression. A similar explanation was offered to explain the apparent existence of pull-apart basins on the transpressional Macquarie Ridge complex (Daczko *et al.* 2003).

Strike-slip bends of the southern Dead Sea fault system

The dominant bend structures in this slip-parallel southern segment of the fault are an array of three large rhomboidal pull-apart basins in the Gulf of Aqaba (Elat) (Ben-Avraham *et al.* 1979; Garfunkel & Ben-Avraham 1986; Ben-Avraham & Tibor 1993; Fig. 21). These basins underlie the 2-km-deep Gulf of Aqaba (Elat) that is flanked by

1-km-high mountains. Recent work by Ehrhardt *et al.* (2005) at the northern end of the northernmost basin show a gentle lazy-Z-type fault geometry rather than the more rectilinear fault geometry proposed by previous workers. To the north, the Dead Sea fault trace forms a linear scarp in the rift-like Arava Valley and intersects the Dead Sea rhomboidal pull-apart basin in the north.

With a 150 km length and an 8–10 km width, the Dead Sea Basin is the largest pull-apart on the Dead Sea fault system (ten Brink *et al.* 1999; Fig. 21). The basin started to form about 15 Ma or earlier, during the first phase of strike-slip motion, and reached about half its present length before the end of the Miocene (Garfunkel & Ben-Avraham 1996). Its sediment fill, including a massive salt layer, is about 5 km over half its length and reaches a maximum of about 10 km. According to mapping by Ben-Avraham *et al.* (1990) and Garfunkel & Ben-Avraham (1996), the pull-apart grew by becoming longer in the direction parallel to strike-slip motion, as predicted in the standard pull-apart model (Fig. 7). More recent subsurface mapping by Lazar *et al.* (2006) has emphasized basin opening by rotation and opening along a series of blocks underlying the basin. The lack of high-quality imaging of the extended basement beneath the Dead Sea pull-apart basin precludes a conclusive statement about the precise mechanism of basin opening.

North of the Dead Sea pull-apart, two other smaller but prominent pull-aparts include the Sea of Galilee, where motion is now thought to occur only on one master fault (Hurwitz *et al.* 2002) and the Hula pull-apart basin (Heimann & Ron 1987; Zilberman *et al.* 2000) that is adjacent to the Lebanese restraining bend.

Lebanese gentle restraining bend

The lazy-Z gentle restraining bend is cross-cut by the northern continuation of splayed faults of the Dead Sea fault system including a thrust splay extending along the Mediterranean coast of Lebanon (Daeron *et al.* 2005; Gomez *et al.* 2007) (Fig. 22b). Of these splays, only one fault, the Yam-mouneh, appears to be through-going and interconnects the northern and southern Dead Sea fault zone. Gomez *et al.* (2006, 2007) have proposed that transpressive deformation is partitioned into 4.0 mm/a of strike-slip movement along the Yam-mouneh strike-slip fault, and 2.8 mm/a of orthogonal shortening on folds affecting the coastal area of Lebanon. Schattner *et al.* (2006) has mapped offshore convergent structures which are continuous with the onland shortening structures. Evidence for large-scale block rotations from

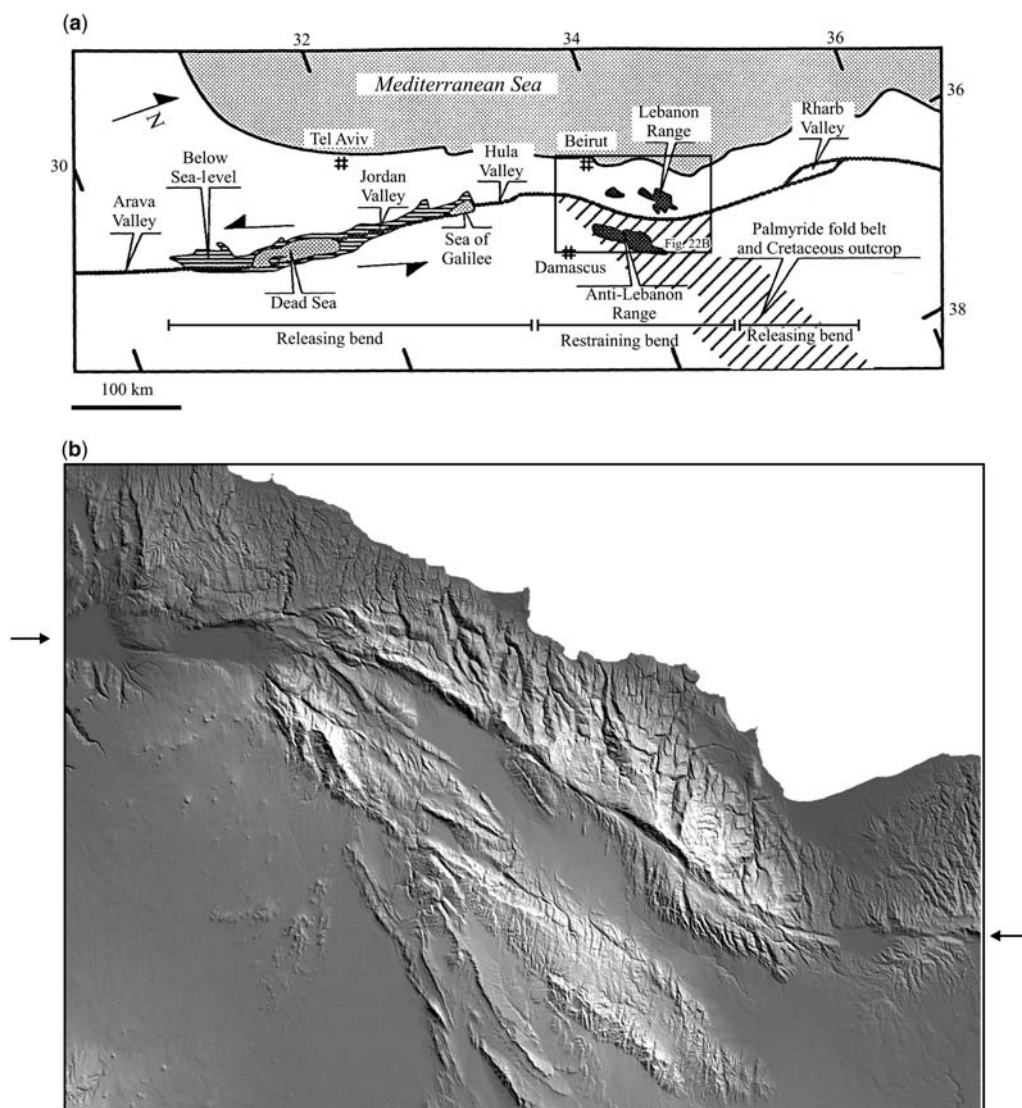


Fig. 22. (a) Schematic map of the active traces of the Dead Sea fault system in the Middle East, from a more detailed map in Figure 21. An inferred paired restraining- and releasing bend is indicated. Unlike the restraining bends of the San Andreas fault system, shown in Figures 16–20, the Lebanon restraining bend probably formed by inversion of an obliquely intersecting rift system of Cretaceous age (the present-day inverted Palmyride fold belt). See the regional tectonic map in Figure 21 for location. (b) SRTM topographic image of the Lebanon Range and Anti-Lebanon Range from Daeron (2005). Arrows indicate the main trace of the fault.

palaeomagnetic data is interpreted by Gomez *et al.* (2007) as an earlier phase of motion prior to the formation of a single throughgoing strike-slip fault zone in a manner similar to that described by Wesnousky for the Walker Lane fault zone in the western US.

Strike-slip bends of the northern Dead Sea fault system

The Ghab rhomboidal pull-apart basin is interpreted by Brew *et al.* (2001) as having formed during the second phase of Dead Sea fault zone evolution,

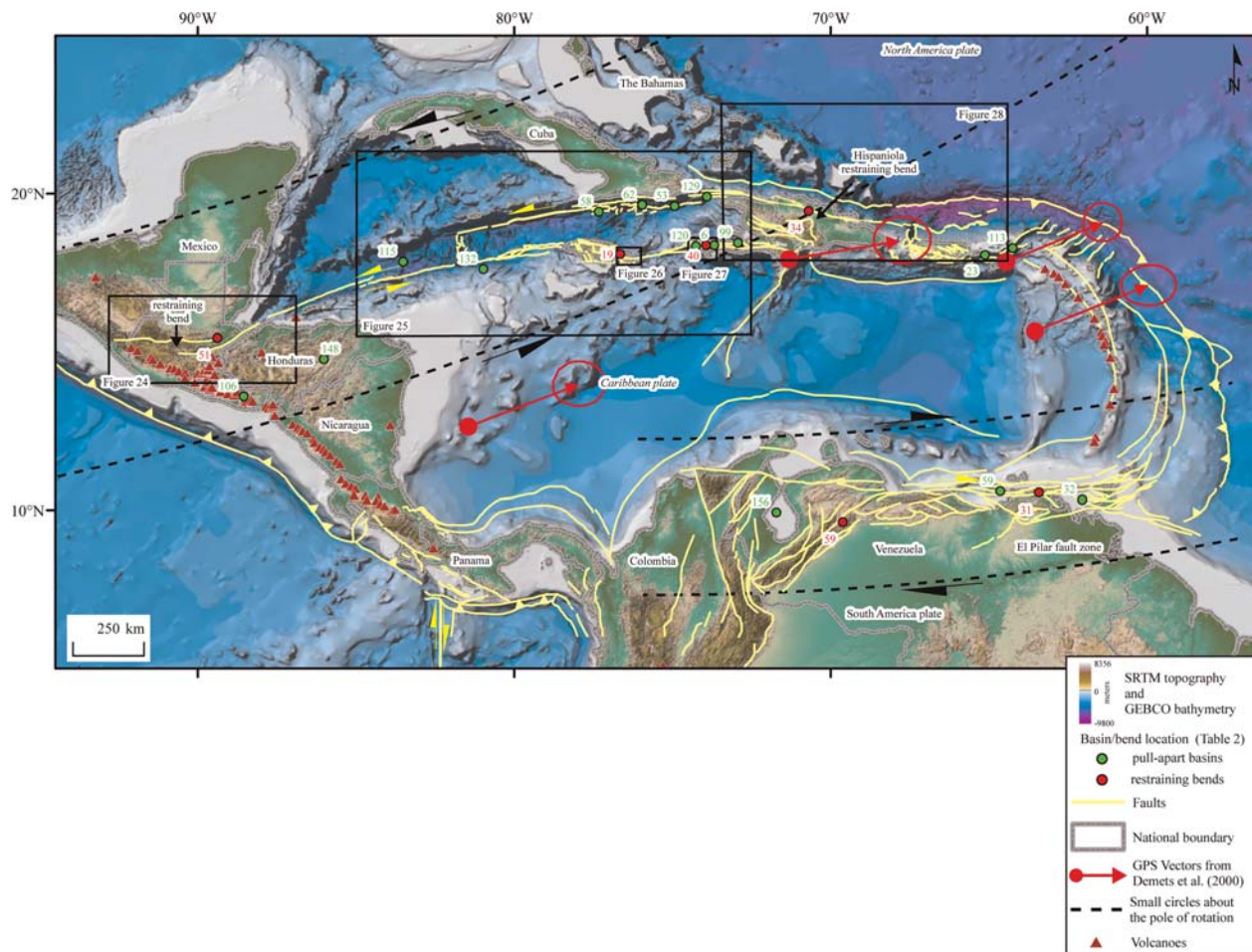


Fig. 23. For caption see p.77.

starting in the latest Miocene and earliest Pliocene. Some extension and subsidence has occurred on transverse normal faults, but most subsidence has occurred along the eastern master fault. Gomez *et al.* (2005) has pointed out that the GPS-based poles of rotation predict that the northern Dead Sea fault segment containing the Ghab pull-apart is highly transpressive, with the angle of intersection between the fault trace and the small circles being as much as 25° . This transpression can explain the uplifted margins of the northern Dead Sea fault that extend well beyond the extensional influence of the Ghab pull-apart basin.

Paired-bend interpretation. The geometry of the Lebanese gentle restraining bend with topographic elevations over 2 km and the Dead Sea–Jordan Valley with minimum elevations of -400 m along the Dead Sea fault system suggests that the two structures constitute a large paired bend (Fig. 22a). As in the cases of the Californian paired bends, there is a similarity in the width and dimensions of the adjacent bends that may be explained by a mass balance between the excess crustal mass in the restraining bend with a thinned crustal mass in the Dead Sea pull-apart area.

Early paired-bend history. Seismic reflection and drill-hole data from Syria suggest that the Lebanese restraining bend formed at the site of a Permian–Triassic (or even older) failed rift of the Levantine margin that was inverted to form the Lebanon and Anti Lebanon Mountains and Palmyrides fold–thrust belt by Late Mesozoic and Cenozoic compression (McBride *et al.* 1990; Fig. 22a). Some of the inversion of the Palmyrides may be a far-field effect of closure or ‘zippering’ between Arabia and the Zagros collisional belt (Hempton 1987; Brew *et al.* 2001). The ancient rift trend subtends an angle of about 45° with the Dead Sea fault system. Initial formation of a paired bend is inferred to predate the formation of the Dead Sea–Jordan Valley releasing bend as a result of the fault re-adjusting to its bent configuration (Fig. 12c) (Mann *et al.* 2007a). Unfortunately, the initial age of the Lebanese restraining bend is not well known.

The timing of the Dead Sea pull-apart is well constrained. Ten Brink and Ben-Avraham (1989) interpret the localized occurrence of evaporites to

indicate that the Dead Sea pull-apart formed in Middle to Late Pliocene times. Farther south, Ben-Avraham (1985) has inferred that the Gulf of Elat pull-aparts are also Pliocene in age (Fig. 21).

Tectonic setting and geological development of bends along the northern and southern Caribbean strike-slip fault systems

Caribbean strike-slip systems. The Caribbean plate is moving eastward relative to North and South America along two strike-slip fault systems along its northern and southern edges (Fig. 23). As the Caribbean plate moves eastward, the tips of these strike-slip faults also propagate eastward as the edge of the Lesser Antilles subduction zone. For this reason, these and tectonically similar faults in other parts of the world have been recently classified as *STEP faults*, or *subduction–transform edge propagators* (Govers & Wortel 2005).

The North America–Caribbean strike-slip system is a broad zone of deformation, extending 3200 km from northern Central America in the west to the Lesser Antilles Arc in the east. Most of the plate-boundary zone faults are submarine, with sub-aerial exposures of faults in northern Central America (Pflaker 1976; Fig. 24), in Jamaica and the southern peninsula of Hispaniola (Haiti) (Fig. 27) and in north-central Hispaniola (Fig. 28). The dominant structural element in the central plate boundary zone is the Cayman Trough, a submarine pull-apart basin formed by at least 1100 km of oceanic spreading at the Mid-Cayman spreading centre (Rosencrantz *et al.* 1988; Leroy *et al.* 2000; Fig. 25b). The spreading centre has been active since the Mid-Eocene, and is currently spreading at a rate of 15 mm/a (Rosencrantz *et al.* 1988).

The direction and rate of relative motion of the Caribbean plate has been quantified by GPS measurements (DeMets *et al.* 2000; Jansma *et al.* 2000; Mann *et al.* 2002), and these data can be used to predict the style of deformation along the length of both the northern and southern Caribbean

Fig. 23. Regional tectonic map of the northern and southern Caribbean strike-slip fault systems, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilations by Mann *et al.* (2002), Audemard & Serrano (2001) and Cowan *et al.* (2000). Black lines in the northern Caribbean represent small circles of rotation about the North America–Caribbean pole, based on GPS results from DeMets (2001). Dashed black lines in the southern Caribbean represent small circles of rotation about the South America–Caribbean pole, based on GPS results from Perez *et al.* (2001) and Weber *et al.* (2001).

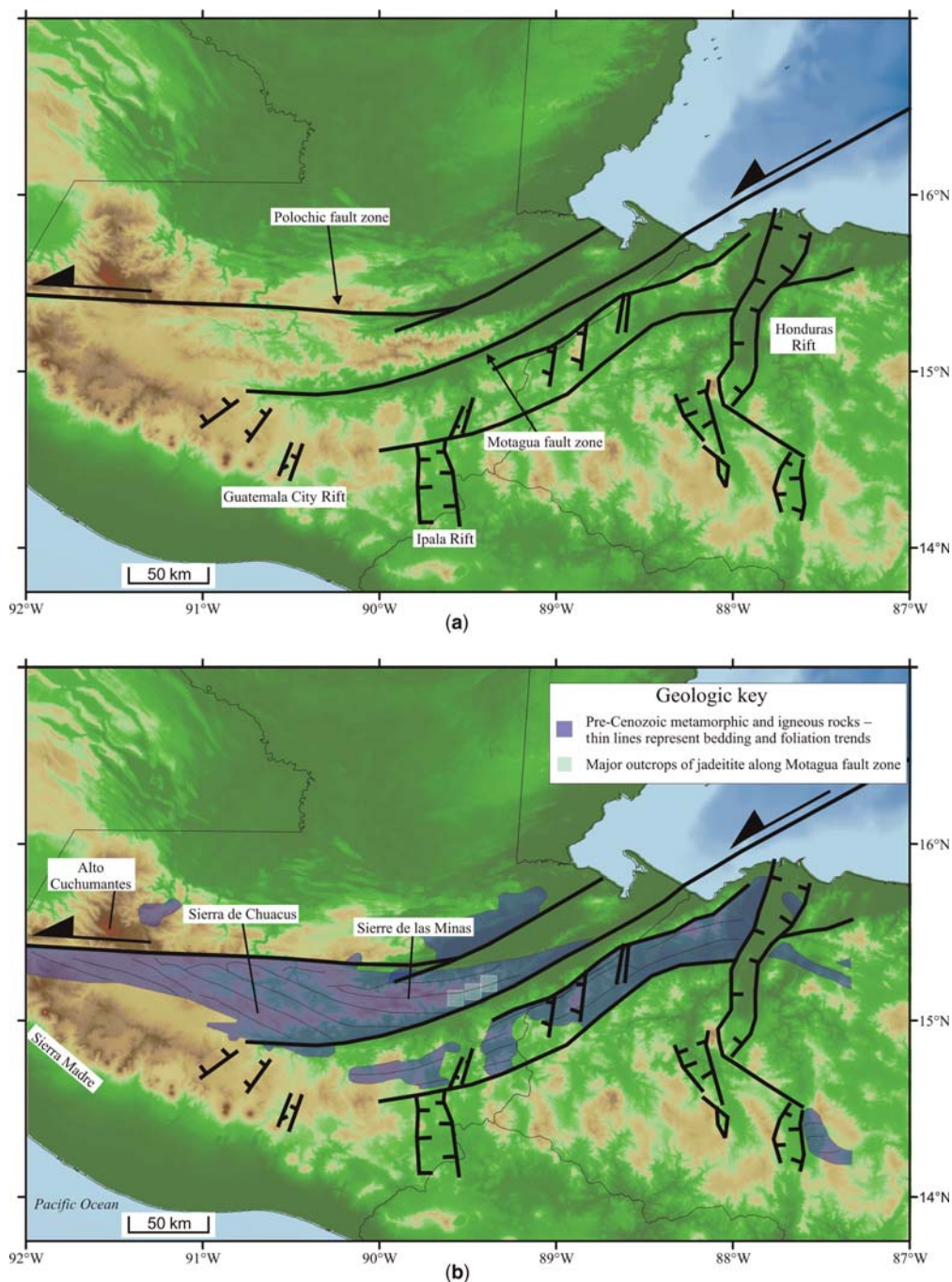


Fig. 24. (a) Topography of northern Central America and its relation to the left-lateral strike-slip faults separating the North America and Caribbean plates, modified from Mann & Gordon (1996). Note that the direction of relative plate motion of North America and the Caribbean is approximately east–west, and that the areas of highest and lowest topography are adjacent to the segment of the fault most oblique to that direction. The structure is interpreted as a paired bend – with a restraining bend in the mountainous area to the west, and a releasing bend in the deep, fault-controlled valleys and rifts to the east. (b) Map showing the ages of exposed rock units in northern Central America. Mann *et al.* (1996) proposed that the topographic highlands and exposures of deeper crustal rocks in the Sierra de las Minas, Sierra de Chuacus, and the eastern part of the Sierra Madre may be related to the combined effects of restraining bends in that area. Blueschist (jadeitite) outcrops are restricted to the area of the three squares shown in the Sierra de las Minas.

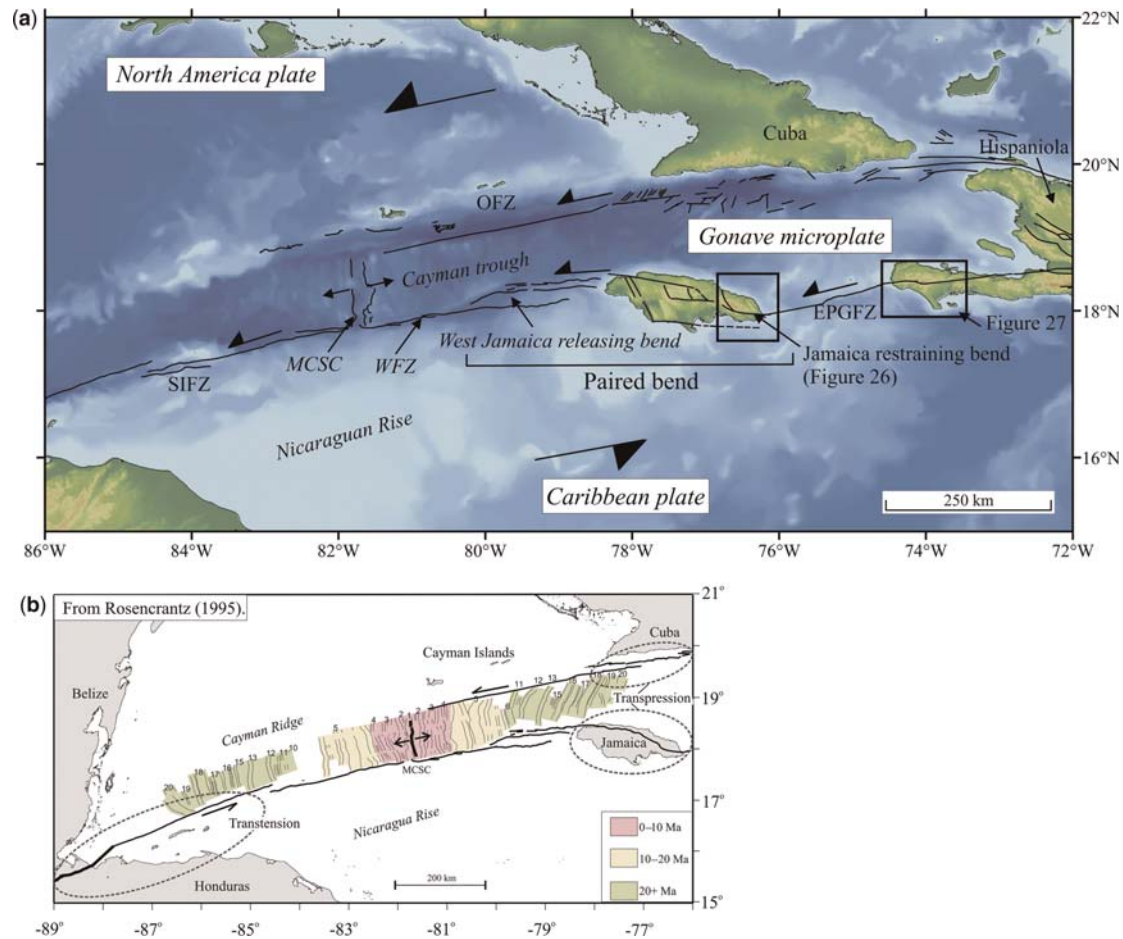


Fig. 25. (a) Regional tectonic map of the North America–Caribbean strike-slip fault system, using SRTM topography and GEBCO bathymetry as a base map. Black lines are active faults. An inferred paired restraining- and releasing bend is indicated. (b) Map of oceanic crust in the Cayman Trough, with black lines representing magnetic anomalies formed at the Mid-Cayman spreading centre (MCSC), as identified by Rosencrantz (1995). The oldest magnetic anomalies of Early Eocene age (c. 43 Ma) indicate that the Cayman Trough has accumulated about 1000 km of slip during its long life.

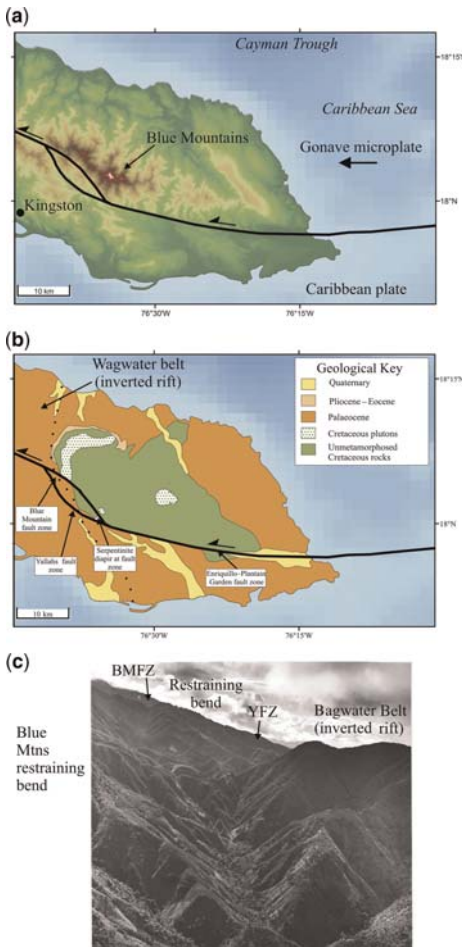


Fig. 26. (a) SRTM topography of eastern Jamaica, and its relation to active traces of the left-lateral strike-slip faults separating the Gonave microplate in the north (see Fig. 25a for location) from the Caribbean plate to the south. Note that the direction of North America–Caribbean relative motion is east–west, and that the area of higher topography is adjacent to the segment of the fault most oblique to this direction. (b) Map showing the ages of exposed rock units in the restraining-bend area. Note the presence of Cretaceous amphibolites, greenschists and blueschists along the upthrown side of the restraining bend, and fan deltas composed of Quaternary gravels that form a radial pattern around the restraining-bend uplift. The Palaeogene Wagwater rift (inverted during the Late Miocene to Recent) is inferred to be the basement structure responsible for the formation of the restraining bend. (c) Aerial view looking SE along the Blue Mountain (BMFZ) and Yallahs (YFZ) fault zones that form reverse faults bounding the SE edge of the Jamaica restraining bend and Blue Mountains. The Yallahs Fault originated as a Palaeogene, synsedimentary normal fault bounding the now-inverted Wagwater Rift.

strike-slip fault zones (Fig. 23). Small circles of rotation between the North America and Caribbean plates predict transtension along the western part of the plate boundary in northern Central America and the Cayman Trough; pure strike-slip near the Cayman spreading centre; and transpression in a belt extending from southern Cuba to Puerto Rico at rates of about 20 mm/a (Mann *et al.* 2003; Fig. 23).

In northern South America, GPS surveys have shown that the eastern part of the Caribbean–South America plate boundary is relatively simple, with small circles parallel to the trend of east–west strike-slip faults (Perez *et al.* 2001; Weber *et al.* 2001; Fig. 23). Rates are about 20 mm/a, and transpression is predicted only in the easternmost area near Trinidad (Weber *et al.* 2001). In this section I will focus on the current state of knowledge of its major restraining and releasing bends along the northern and southern boundaries within the tectonic framework established by these plate-motion studies. More detailed descriptions of the Caribbean plate and its strike-slip fault systems, including their origin, offset history and palaeogeography, are given by Pindell & Barrett (1990) and Mann *et al.* (1995).

Strike-slip bends of the North America–Caribbean strike-slip system

Northern Central America restraining bend. The western end of the main strike-slip fault (Motagua fault zone) of the North America–Caribbean strike-slip plate boundary curves to the NW in northern Central America to form a large gentle bend with high topographic elevations up to 4000 m and elongate, fault-aligned exposures of metamorphic rocks including the high-pressure mineral jadeite (Mann & Gordon 1996; Harlow *et al.* 2004; Fig. 24). This curvature of the fault intersects small circles of plate rotation to predict a strong component of tectonic transpression along the curved part of the fault trace, which ruptured along a 200-km distance to produce the *M7.2* earthquake of 1976 (Pflaker 1976).

Interactions between the Motagua Fault and other neighbouring strike-slip faults may also play a role in the regional deformation and topographic uplift in this highly curved area of the fault. In addition to the gentle bend along the Motagua fault zone, Mann & Gordon (1996) infer a 300-km-long and 100-km-wide sharp restraining bend or stepover between the Motagua and Polochic faults (Fig. 24a). This stepover localizes the extremely steep-sided Sierra de Minas, bounded by the two master faults and reaching elevations of over 3 km (Fig. 24a). Guzman-Speziale & Meneses-Rocha (2000) propose a tectonic model, whereby

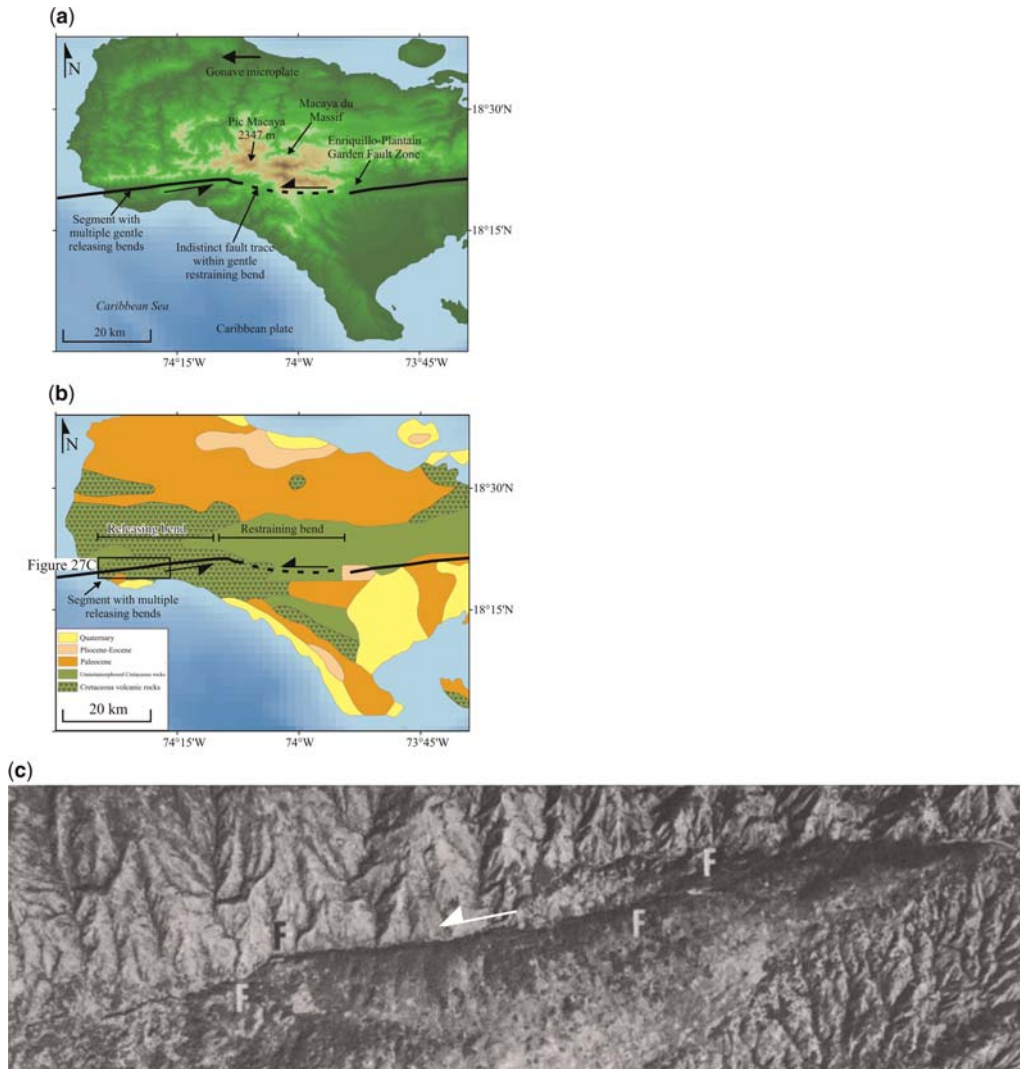


Fig. 27. (a) SRTM topography of the western part of the southern peninsula of Haiti, and its relation to active traces of the left-lateral strike-slip faults separating the Gonave microplate in the north (see Fig. 25a for location) from the Caribbean plate to the south. Note that the direction of North America–Caribbean relative motion is east–west, and that the area of higher topography (Massif du Macaya) is adjacent to the segment of the fault most oblique to this direction. The structure is interpreted as a paired bend, with the Massif du Macaya restraining bend to the east, and a releasing bend in the deep, fault-controlled valleys to the west. (b) Map showing the ages of exposed rock units in the restraining-bend area. Cretaceous volcanic and sedimentary rocks are exposed in a belt parallel to the fault; a domal structure is not as prominent as in eastern Jamaica (as seen in Fig. 26b) probably because the width of the restraining-bend stepover is only 15 km (as opposed to the 55-km distance in Jamaica). (c) Aerial photograph of incipient pull-aparts forming in the releasing-bend part of the paired-bend structure interpreted in Figure 27a and b (from Mann *et al.* 1983; original photo scale: 1:40 000). Letter F marks the approximate termination of dominant oblique-slip fault segments at releasing bends. Overlapping of strike-slip fault segments appears to be occurring at the eastern bend. Basin shapes suggest young pull-aparts similar to those compiled in Figure 3a–f.

motion from the Polochic Fault is transferred to another large, sharp restraining bend that they postulate on the basis of Late Neogene folding and faulting in southern Mexico.

The eastern part of the curved Motagua Fault is characterized by active rifts trending at right angles to the fault zone (Pflaker 1976; Fig. 24a). These features either represent fault-termination structures

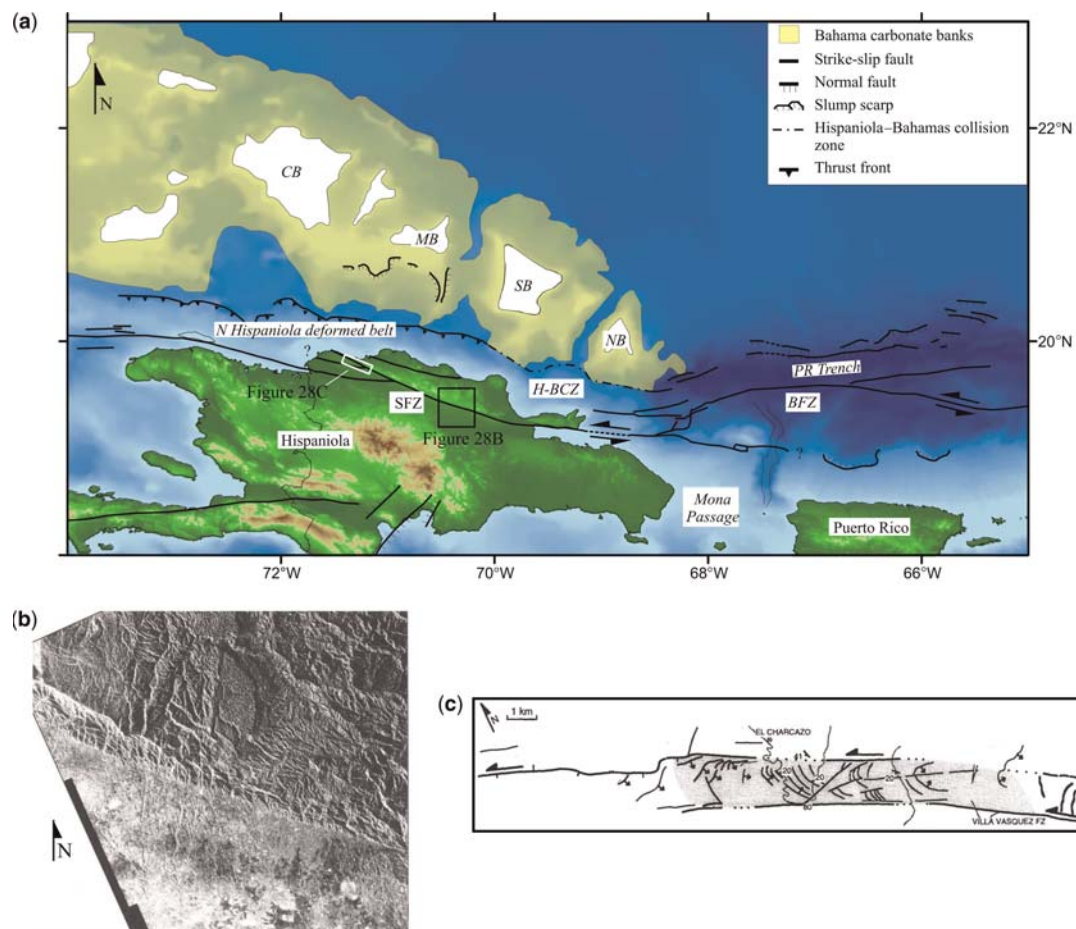


Fig. 28. For caption see p. 83.

(Guzmán-Speziale 2001) similar to those shown in Figure 5, or may represent an extreme example of partitioning of extension along the highly curved and transtensional eastern segment of the Motagua fault zone (Mann *et al.* 2007b; Rogers & Mann 2007).

Cayman Trough extreme pull-apart basin. The Cayman Trough is an extreme rhomboidal pull-apart basin that forms a 100 by 1400 km elongate depression with maximum water depths greater than 6000 m (Fig. 25b). North–south trending ridges in the centre of the trough are produced by slow spreading along the Mid-Cayman spreading centre. The northern and NW margins of the trough are bounded by elongate submarine ridges that extend near or above sea-level. The SE margin of the basin is marked by the island of Jamaica and a prominent bathymetric depression west of Jamaica: the West Jamaica pull-apart basin (Fig. 25a). Studies of marine magnetic anomalies and aeromagnetic anomalies within the Cayman Trough by Rosencrantz (1995) show similar patterns of basin opening that began quickly from about 49 to 26 Ma then slowed down considerably between 26 and 20 Ma, and maintained a slow but steady spreading rate from 20 Ma to the present-day (Fig. 25b). These anomalies show that the basin maintained its original master-fault separation distance of about 100 km through its 49 Ma of evolution. From present-day plate motions, the western and central part of the trough is transtensional to pure strike-slip, whilst the eastern part near Cuba and Jamaica is transpressional (Fig. 23). Transpression in this area has been documented in the form of narrow fold–thrust belts by Calais & Mercier de Lepinay (1991) and Leroy *et al.* (1996).

Paired bends in Jamaica and western Haiti. Jamaica is uplifted along a gentle restraining bend in the eastern part of the island as discussed in detail by Mann *et al.* (2007a; Fig. 26a). Strike-slip faulting occurs along the Enriquillo–Plantain Garden fault zone which bounds the southern edge of the Gonave microplate (DeMets & Wiggins-Grandison 2007; Fig. 25a). Mann *et al.*

(1995, 2003) proposed that the Gonave microplate detached from the North America plate about 5 Ma in response to oblique collision between the Caribbean plate and the Bahama carbonate platform. The direction of Gonave–Caribbean plate motion is roughly east–west at rates of *c.* 10 mm/a (DeMets & Wiggins-Grandison 2007).

In eastern Jamaica the Enriquillo–Plantain Garden Fault forms a prominent topographic limit separating older, exhumed Cretaceous–Palaeocene igneous, metamorphic and sedimentary rocks of the Blue Mountains from Eocene–Late Quaternary clastic and carbonate rocks fringing the domal uplift (Mann *et al.* 1985) (Fig. 26b). Elongate outcrops of blueschists and serpentinites are locally emplaced along the faulted edge of the restraining bend. As in many gentle bends, the topographic relief of the Jamaican bend is greatest near the most curved segments of the strike-slip faults bounding the bend. Other restraining bends are present in the central and western parts of the island (Mann *et al.* 2007a).

As in the case of the gentle Lebanese restraining bend, the Jamaica bend appears to have nucleated on a pre-existing Palaeogene rift structure that trends about 45° to the strike-slip fault (Mann *et al.* 1985, 2007a). This rift is now inverted and forms high mountains adjacent to the bend (Fig. 26c).

The West Jamaica pull-apart basin is inferred to represent the rhomboidal pull-apart basin paired to the restraining bends found in Jamaica (Fig. 25a). The southward step in the fault to form the 2-km-deep depression shifts the fault trace to the original trend of the Enriquillo–Plantain Garden fault zone east of Jamaica in a manner similar to that seen in the California and Lebanese bends.

Southern Haiti paired bend. A smaller paired bend structure is present along the Enriquillo–Plantain Garden fault system on the western end of the southern peninsula of Haiti (Fig. 27a). A 10-km wide right-step in the fault trace produces the 2-km-high Massif de la Macaya which exposes Late Cretaceous limestone and basalt at its core (Fig. 27b). Just to the east of the restraining bend, the fault steps southward along a fault trace

Fig. 28. (a) SRTM topography and GEBCO bathymetry of the islands of Hispaniola and Puerto Rico in the NE Caribbean, and their relationship to active traces of the left-lateral strike-slip faults separating the North America plate in the north from the Caribbean plate and Gonave microplate to the south. The north step in the North America–Caribbean strike-slip fault system across northern Hispaniola is related to the oblique collision and partial subduction of the Bahama carbonate platform on the North America plate with the island of Hispaniola on the Caribbean plate and Gonave microplate. Key to abbreviations: CB, Caicos Bank; MB, Mouchoir Bank; SB, Silver Bank; NB, Navidad Bank; SFZ, Septentrional left-lateral strike-slip fault zone; PR Trench, Puerto Rico Trench. (b) Seasat radar image, showing uplifted topography and exposure of older rock units along the central curved part of the Septentrional fault zone, modified from Mann *et al.* (1998). (c) Sharp restraining bend or push-up structure along the western Septentrional fault zone, modified from Mann *et al.* (1998).

containing four left-stepping gentle releasing bends interpreted to be incipient pull-apart basins (Mann *et al.* 1983; Fig. 27c).

Hispaniola gentle restraining bend. Hispaniola forms a prominent restraining bend in the main trace of the North America–Caribbean strike-slip plate-boundary fault, the Septentrional fault zone (Mann *et al.* 1984; Mann *et al.* 1998; Prentice *et al.* 2003). The bend is gentle, with a maximum curvature of 10° in the Septentrional fault zone (Fig. 28a). The northern range of Hispaniola is on the northern upthrown side of the Septentrional strike-slip fault, and it forms a prominent topographic limit separating older, Cretaceous–Early Pliocene igneous metamorphic (including blueschists), and sedimentary rocks from a basal section in the valley to the south (Fig. 28b). The direction of Caribbean–North America plate motion in this area reaches its maximum amount of transpression in this area, with rates of about 20 mm/a (Mann *et al.* 2002). The correspondence of structural and topographic doming suggests that transpression is active.

The formation of the curvature and gentle bend in the Septentrional fault zone is probably related to the oblique collision of the Bahama carbonate platform on the North America plate (Dolan *et al.* 1998; Mann *et al.* 2003). Based on tectonic reconstructions, Mann *et al.* (1999, 2003) estimate that this oblique collision between the Bahama Platform and the NE Caribbean plate in Hispaniola began about 5 Ma ago and continues to the present day.

The Septentrional fault zone is the site of a single, sharp restraining bend (Fig. 1c) along one of its fault splays. The Villa Vasquez bend measures 15 km long by 3 km and exposes a folded belt of Miocene and Pliocene rocks with a badlands-type topography. The location of this sharp bend within an overall gentle bend suggests that tectonic transpression may have led to its formation.

Strike-slip bends of the South America–Caribbean strike-slip system

Introduction. There are two zones of strike-slip faulting in northern South America that accommodate relative motion between different plates or blocks in different tectonic settings (Fig. 1, Table 1). The simpler system is the El Pilar strike-slip fault system from central Venezuela to Trinidad that accommodates eastward motion of the Caribbean plate relative to South America (Fig. 23). The El Pilar fault zone can be classified as a continental-plate-boundary strike-slip fault system that can be described by small circles of plate motion, as shown on Figure 23 (Woodcock 1986;

Table 1). Two rhomboidal pull-aparts are present along the El Pilar fault system: the Cariaco Basin in Venezuela (Schubert 1982) and the Gulf of Paria Basin in easternmost Venezuela and Trinidad (Flinch *et al.* 1999; Babb & Mann 1999) along with one intervening restraining bend on the Araya–Paria Peninsula (Audemard *et al.* 2006).

Tectonic setting and geological development of bends along the Scotia Sea strike-slip fault systems

Scotian strike-slip systems. The Scotia Sea region – bounded by an arc to the east and strike-slip fault systems to the north and south – has long invited tectonic comparisons to the Caribbean plate (cf. Royden 1993) (cf. Figs 23 & 29). However, the plate structure, active plate kinematics and tectonic history of the two plates differ in many respects. The Scotia region consists of two small plates: the larger Scotia plate to the west largely composed of oceanic crust, and the small South Scotia plate to the east formed in a back-arc position west of the South Sandwich Arc (Fig. 29). Both plates have formed over the past 30 Ma in response to changing relative motion between the much larger South America and Antarctica plates (Cunningham *et al.* 1995, 1998) and in the case of the smaller South Sandwich plate, to slab rollback of the subducted South Atlantic slab beneath it (Royden 1993).

GPS studies (Smalley *et al.* 2003) and earthquake studies (Pelayo & Wiens 1989; Giner-Robles *et al.* 2003) and plate models (Thomas *et al.* 2003) show that the Scotia plate is bounded to the north and south by left-lateral strike-slip fault systems which deform much of the plate within a left-lateral shear couple. Unlike the Caribbean plate, which experiences a component of north–south shortening between North and South America, the Scotia plate is bounded by two larger plates that are moving in a direction close to east–west, and therefore lack a significant component of north–south shortening (Thomas *et al.* 2003).

Bends along the North Scotia strike-slip fault system. The North Scotia Ridge is a series of islands and submarine banks that forms the northern edge of the Scotia plate and extends 2000 km from the southern tip of South America (Tierra del Fuego) to South Georgia Island in the western South Atlantic Ocean. Like the northern Caribbean, most of the strike-slip system is covered by water, and must be studied using GPS instruments placed on islands, earthquake seismology, or marine geophysical techniques. Earthquake focal mechanisms indicate that motion is east–west strike-slip and that the component of convergence increases in an

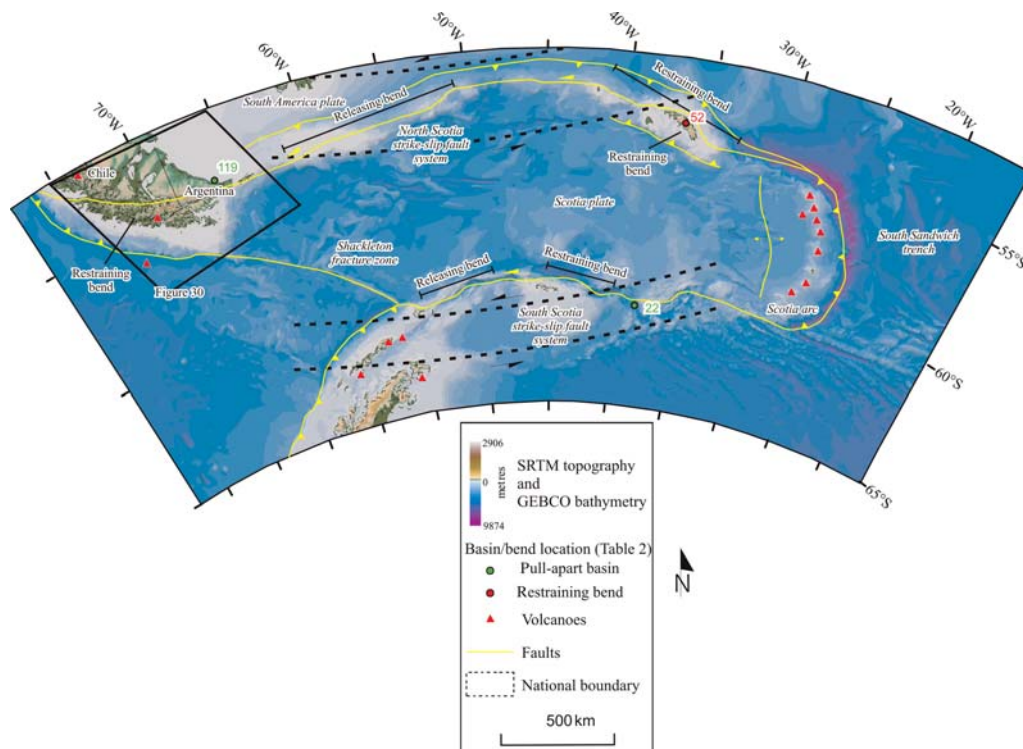


Fig. 29. Regional tectonic map of the north Scotia strike-slip fault system in the area of southernmost South America (Tierra del Fuego), using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilations by Cunningham (1995), Lodolo *et al.* (2003), Mann *et al.* (2003) and Smalley *et al.* (2003). Dashed black lines are small circles about the pole of rotation, from Thomas *et al.* (2003). Note that the direction of Scotia–South America relative motion is east–west, and that the area of higher topography in Tierra del Fuego and South Georgia Island is adjacent to the segment of the fault most oblique to this direction. This large restraining bend is similar to those found in other strike-slip to subduction transition areas, including northern Central America (Fig. 24a & b) and Hispaniola (Fig. 28a).

eastward direction, with a maximum of convergence near South Georgia Island (Pelayo & Wiens 1989; Cunningham *et al.* 1998; Fig. 29). The North Scotia Ridge is flanked to the north by a large accretionary prism believed to have been formed by an oblique component of convergence across the left-lateral strike-slip plate boundary (Cunningham *et al.* 1998). The exact trace of the main strike-slip fault or faults parallel to the North Scotia Ridge is not well mapped, but is inferred on the basis of isolated structural features and bathymetry by Cunningham *et al.* (1998) and Mann *et al.* (2002). The fault makes landfall on Tierra del Fuego and has been mapped as a curving trace by Lodolo *et al.* (2003) (Fig. 30).

Bends have been best studied in Tierra del Fuego where Cunningham (1993) has interpreted the large curving trace of the fault as a gentle restraining bend (Fig. 1c), similar to that described above at the western end of the North America–Caribbean

strike-slip boundary in northern Central America (Fig. 24). Smalley *et al.* (2003) has used GPS to show that plate motion in the bend area of Tierra del Fuego is almost entirely east–west and therefore consistent with large amounts of shortening in the NW-trending part of the bend. This gentle bend explains the formation of elevated topography along the trace of the fault, exposure of high-grade metamorphic rocks, and the deformation of younger sedimentary rocks (Cunningham 1995). Near the SE edge of the bend, Lodolo *et al.* (2003) have interpreted a rhomboidal pull-apart basin at a left-step in the fault trace at Lago Fagnano.

To the west, Vogt *et al.* (1976) and Mann *et al.* (2003) have interpreted South Georgia Island as a major gentle restraining bend formed at a major stepover in the strike-slip fault trace along the North Scotia Ridge. The island trends at right angles to the small circles of plate rotation; is associated with thrust type focal mechanisms;

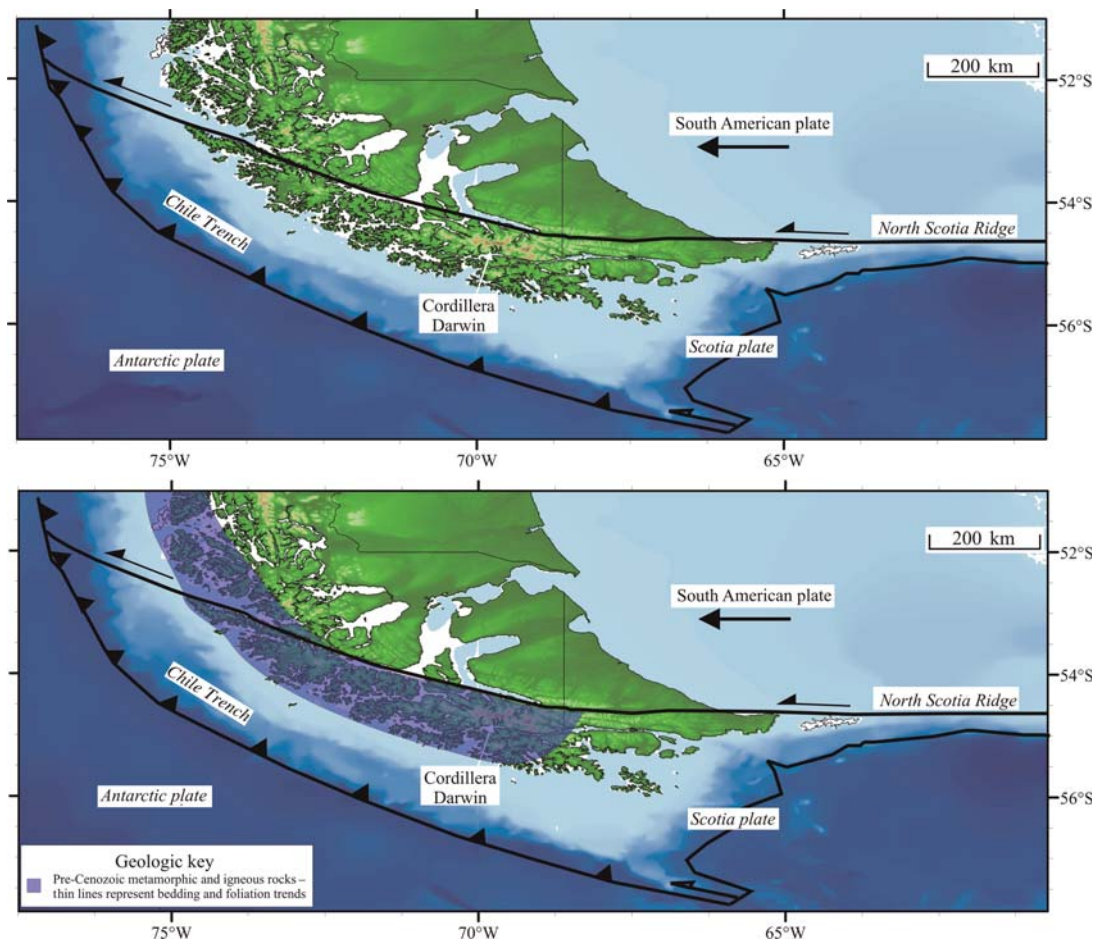


Fig. 30. For caption see p. 87.

mainly exposes highly deformed, older Mesozoic rocks; and is remarkably small (3500 km²) for its maximum elevations over 3 km. Mann *et al.* (2003) have interpreted the South Georgia bend as a result of the indentation of the North Scotia Ridge by the Northeast Georgia Rise, a Mesozoic large igneous province formed during the early opening of the South Atlantic.

Given the curvature of the North Scotia Ridge, it is possible that the South Georgia restraining bend is adjacent to a releasing bend along the poorly mapped western part of the ridge. This hypothesis could be tested by more detailed mapping of the fault trace along the western part of the North Scotia Ridge.

A much larger gentle bend on the Scotia–South America plate boundary is formed by the northward curvature of the plate boundary fault adjacent to the Cordillera Darwin in Tierra del Fuego (Fig. 30). Cunningham *et al.* (1995) attribute the uplift and high topography of there pre-Cenozoic metamorphic and igneous rocks exposed in the Cordillera Darwin to shortening at the bend.

Bends along the South Scotia strike-slip fault system. Left-lateral shear along the South Scotia Ridge occurs on a much more complex and discontinuous morphological boundary than the North Scotia Ridge (Fig. 29). Marine surveys by Acosta & Uchupi (1996) and Galindo-Zaldivar *et al.* (1996) showed a rotating block structure with local deeps that were inferred to be collapsed block edges or pull-apart basins. Earthquakes indicate that much of the margin is transtensional in nature (Thomas *et al.* 2003).

Tectonic setting and geological development of bends along the Alpine–Macquarie Ridge strike-slip fault systems

The 650-km-long right-lateral Alpine fault zone of New Zealand links the subduction zone at the southern end of the Kermadec Trench off the NE coast of the North Island of New Zealand with a smaller subduction zone along the SW end (Fiordland) of the South Island and on the northern Macquarie Ridge complex (Puysegur Ridge) south

of New Zealand (Figs 6a, 31). Oblique subduction beneath the North Island gives way to an increasing component of strike-slip along the Alpine fault zone. The rate of motion between the Pacific and Australia plates varies from 40 mm/a on the North Island to 35 mm/a on the South Island (Fig. 31). For complete reviews of the geology of the Alpine Fault–Macquarie Ridge, the reader is referred to Berryman *et al.* (1992), Lebrun *et al.* (2000) and Meckel *et al.* (2005).

Bends along the northern Alpine fault zone

The Alpine Fault is remarkably linear over much of its length. At its northern end, the fault splays into five major, parallel plays that are collectively called the Marlborough faults (Fig. 32a). The northernmost splay, the Wairau Fault, shows a gentle restraining bend that has been compared with the Transverse Ranges (Yeats & Berryman 1987) but is not associated with marked topographic uplift (Berryman *et al.* 1992).

More striking are the fault splays which curve in a more easterly direction that is transtensional with respect to the small circles of plate rotation (Fig. 32a). This small transtensional environment related to the Marlborough fault splays is the only tectonic environment for pull-aparts on the onland Alpine fault zone.

These faults provide the active tectonic setting for the spindle-shaped Poplars Grove and Glynnwy Lake pull-apart basins (Freund 1971; Cowan 1990) and the much larger, rhomboidal Hanmer Basin (Wood *et al.* 1994). The misalignment of the southern edge of the Hanmer Basin with the regional slip vector has resulted in a local belt of transpression along the SE edge of the basin.

Central Alpine fault. The Central Alpine fault onshore is a straight trace that lacks major bends (Berryman *et al.* 1992; Fig. 31). The small circles of rotation intersect the fault trace at a small angle that are responsible for some small, shallow-crustal transpressional structures not visible at the scale of Figure 31 (Norris & Cooper 1995). This segment of the fault is characterized by a 50-km-long area of higher uplift rate, more relief, deeper exhumation,

Fig. 30. (a) Regional tectonic map of the North and South Scotia strike-slip fault systems, using SRTM topography and GEBCO bathymetry as a base map. GPS studies show that the direction of plate motion is east–west and left-lateral for both fault systems. Laterally extensive paired bends are interpreted along both boundaries. (b) Regional tectonic map of the North and South Scotia strike-slip fault systems, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilations by Mann *et al.* (2003). GPS studies show that the direction of plate motion is east–west and left-lateral for both fault systems. Laterally extensive paired bends are interpreted along both boundaries. Metamorphic outcrop (from Cunningham 1991) is shown in purple.

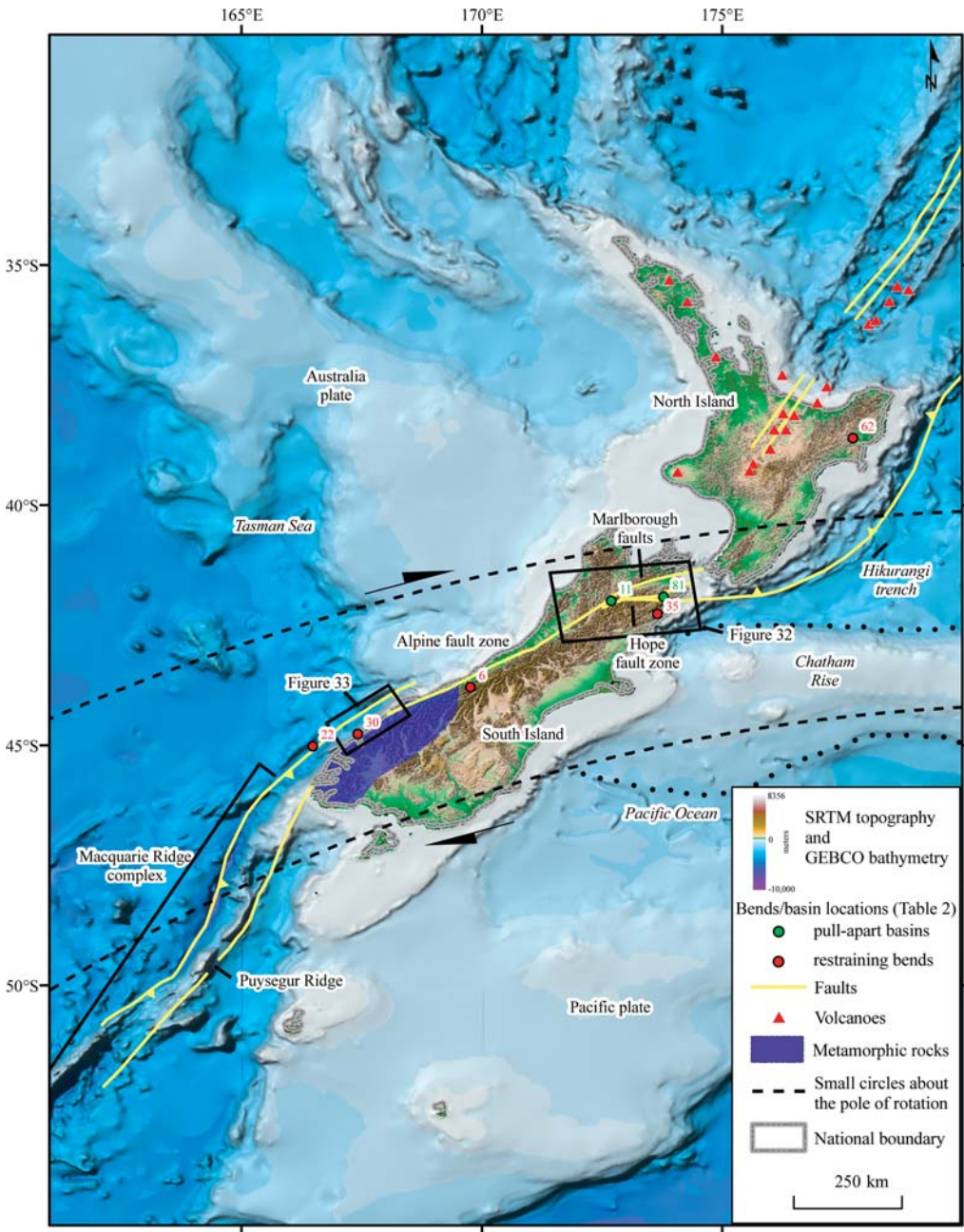


Fig. 31. Regional tectonic map of the Alpine strike-slip fault system in the area of New Zealand and the northern Macquarie Ridge complex, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults, taken from a compilation by Barnes *et al.* (2005). Dashed black lines represent small circles of rotation.

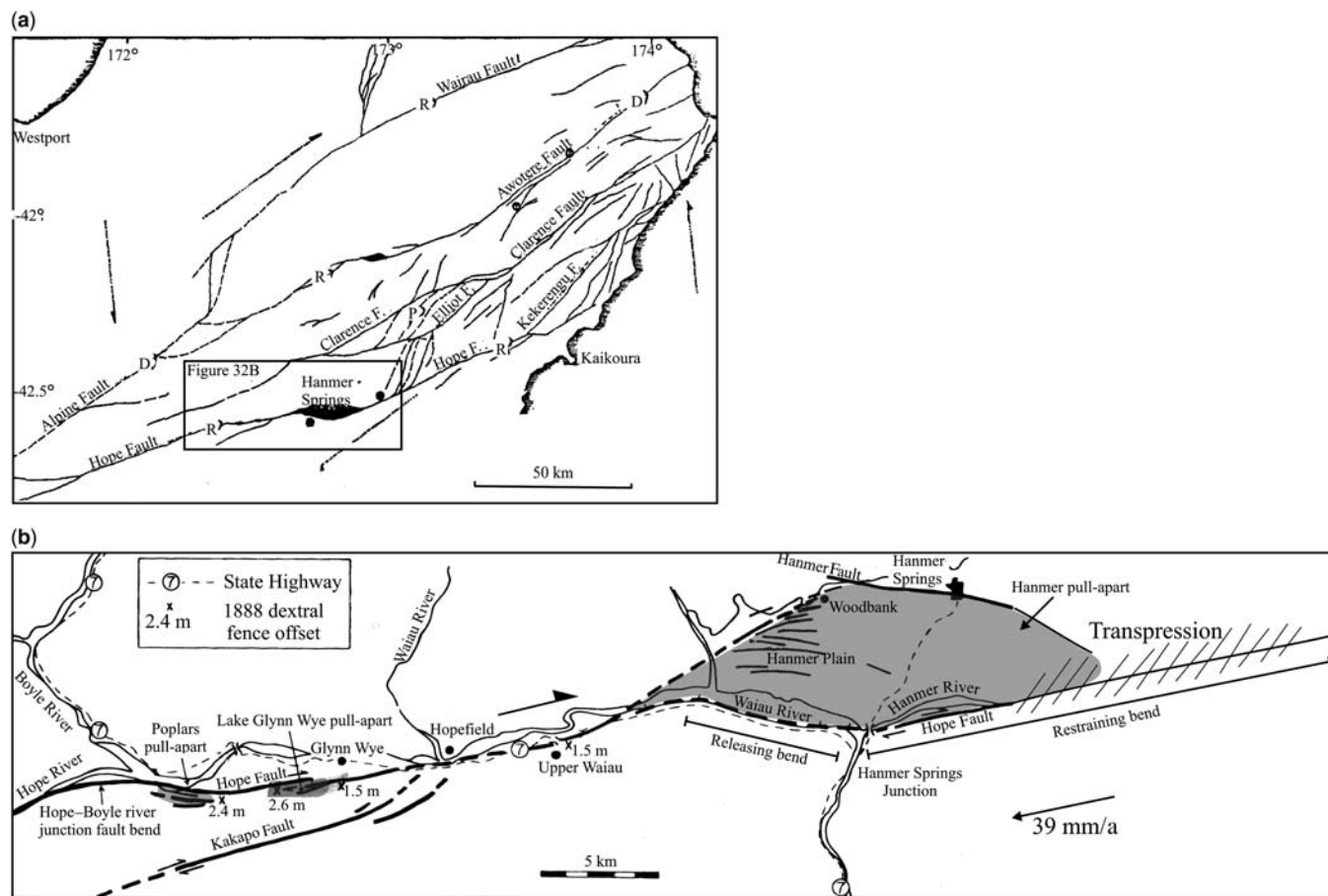


Fig. 32. (a) Map of the Marlborough strike-slip fault system at the northern end of the South Island, from Freund (1974). Black dots show measured offsets on the faults. Black areas are the larger pull-aparts developed on the Hope and Atwater fault zones. Note that at this scale the faults show no obvious paired-bend geometries. Instead, the subparallel strike-slip faults are uniformly right-stepping in a NE direction. (b) More detailed map of the Hope fault zone, showing the array of three pull-apart basins: Poplars, Lake Glynn Wye and Hanmer (modified from Wood *et al.* 1994). At this basin scale, these authors inferred one paired zone of a releasing bend and a restraining bend along the southern edge of the Hanmer pull-apart basin.

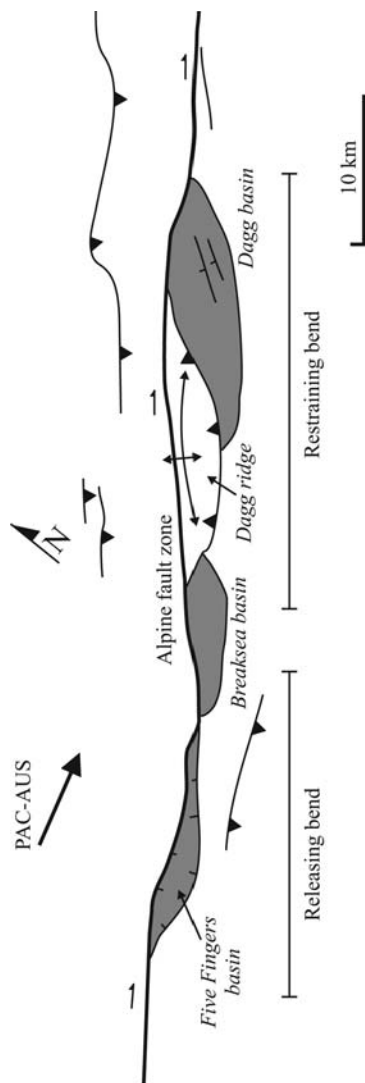


Fig. 33. Schematic map of strike-slip faulting along the west coast of the South Island, by Barnes *et al.* (2005) (cf. Fig. 31 for location), where Pacific–Australia plate motion is slightly oblique to the fault and partitioned into strike-slip faulting and a zone of offshore thrust faulting similar to northern Hispaniola (cf. Fig. 28a–d). A paired bend consisting of a releasing bend (Five Fingers Basin) and a restraining bend (Dagg Ridge) is inferred from the fault pattern.

and a narrower fault width than surrounding areas (Little *et al.* 2005). The hanging-wall uplift of deep-crustal rocks from beneath the Southern Alps in this area occurs as a localized and narrow transpressional uplift (Fig. 1c) that is not associated with a distinct bend in the fault trace.

Southern Alpine fault. The Southern Alpine fault passes offshore and has been surveyed by Barnes *et al.* (2001, 2005) using geophysical methods (Fig. 33). This offshore segment of the fault is characterized by an adjacent releasing bend (Five Fingers Basin) and restraining bend (Dagg Ridge) that together constitute a paired bend (Fig. 33). Barnes *et al.* (2001, 2005) discuss the ephemeral character of the ridges and basins during progressive strike-slip deformation.

Bends along the Puysegur Ridge. The Puysegur Ridge and trench have been shown by marine surveys to exhibit a small subduction feature at the base of its slope associated with a short subducted slab. A zone of strike-slip faults on the trench slope and top of the ridge also accommodate oblique plate motions (Lebrun *et al.* 2000; Fig. 33). The overall morphology of the Puysegur Ridge is that of a lazy-S gentle restraining bend (Fig. 31).

Bends along the Macquarie and Hjort trenches. Transpressional motion persists along the Macquarie and Hjort ridges south of the Puysegur Ridge. The two ridges are separated at a change in their trends by a bathymetric break and deep area in the ridge that is underlain by a single small rhomboidal pull-apart basin (Massell *et al.* 2000; Daczko *et al.* 2003; Fig. 34). Like the central Alpine fault zone, the two ridges exhibit straight morphologies suggestive of transpressional uplifts (Fig. 1c). Daczko *et al.* (2003) postulate other smaller pull-apart basins along the length of the ridges.

The Hjort Trench assumes a more transpressional setting as it curves to the southeast (Fig. 34a). Meckel *et al.* (2005) propose that it is evolving into a subduction zone similar to that observed for the Puysegur Trench.

Tectonic setting and description of natural bends on indent-linked strike-slip faults in active zones of continental collision and tectonic escape

Regions of continental collision and tectonic escape

As compiled by Woodcock (1986), an important tectonic setting for strike-slip faults are areas of continental collision (Fig. 1 & Table 1). Colliding continental

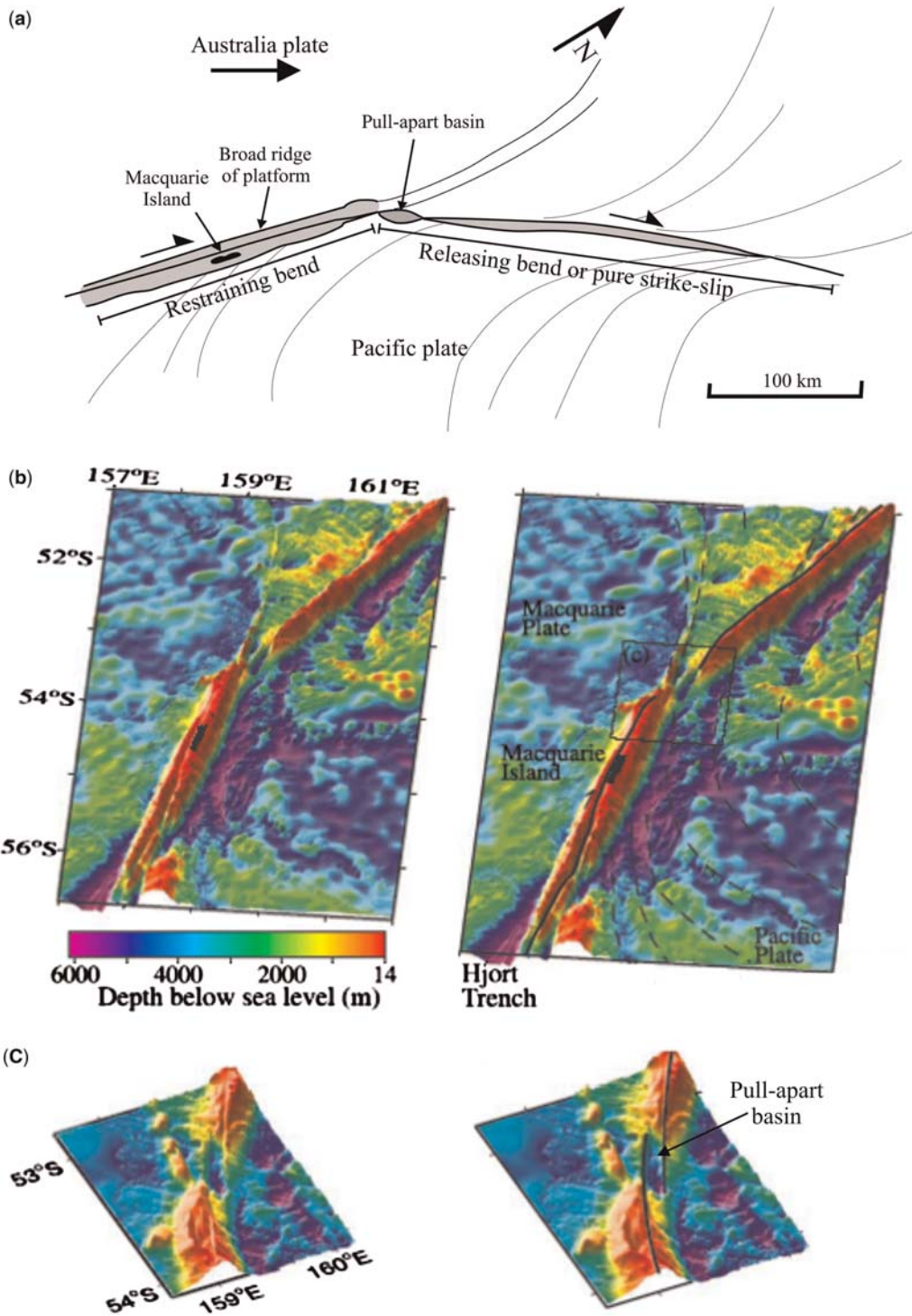


Fig. 34. (a) Schematic map of the Macquarie strike-slip fault system on the Macquarie Ridge south of New Zealand, modified from Daczko *et al.* (2003). The wider platform on which Macquarie Island is located (Macquarie Ridge) is oblique to plate motion and is interpreted to be a restraining bend. Adjacent to the restraining bend is a small but 1.2-km-deep pull-apart basin. The narrower MacDougall Ridge to the north is more parallel to plate motion, and is interpreted as a strike-slip-controlled ridge with less of a transpressional component. Curving fracture zone trends on both sides of the fault are consistent with right-lateral shearing. (b) Multi-beam bathymetric maps of the transition between the two fault segments separated by the deep pull-apart basin, from Daczko *et al.* (2003). (c) Close-up of pull-apart basin, from Daczko *et al.* (2003).

edges commonly have irregularly shaped edges that are mismatched at points of collision and therefore lead to indentation of one continent by the promontory of another. Continental shortening can be taken up thrust faulting and thickening but also by conjugate strike-slip faulting and continental escape of continental fragments along strike-slip faults into uncolided segments of the collision zone. These type of collision-related or 'indent linked' strike-slip faults commonly are associated with releasing fault bends and other types of strike-slip, partly because the escaping blocks undergo significant rotations (Mann & Gordon 1996) and because of reactivation of previous structural grains in the crust (Cunningham 2007). In this section I will review restraining bends in seven areas involving indent-linked strike-slip faults: NW South America, Anatolia, Northern Europe, Iran, Tibet, Mongolia and the Lake Baikal region of Russia.

Tectonic setting and geological development of bends along the active NW South America strike-slip fault systems

The NW corner of South America is a complex, indent-linked zone that accommodates the northward and NE motion or 'escape' of blocks of NW South America into southern South America (Trenkamp *et al.* 2002; Fig. 35a). These indent-linked strike-slip faults are less-continuous, more arcuate and it is more difficult to relate their motion to poles of rotation than the El Pilar strike-slip system to the east that is controlled by a single continental-plate pair (Perez *et al.* 2001; Weber *et al.* 2001; Fig. 23).

Specific strike-slip faults within this system include the Bocono Fault of western Venezuela (Schubert 1980; Fig. 35b), the Santa Marta fault zone of Colombia (Mann & Burke 1984), the Eastern Andean Fault of Venezuela and Colombia (Pennington *et al.* 1979), the Algeciras Fault of Colombia (Velandia *et al.* 2005) and faults of a large restraining bend proposed in the Ecuadorian Andes (Ego *et al.* 1996). GPS shows that all of these faults act across a broad zone to move continental and oceanic fragments in a northward direction, perhaps as a consequence of the collision of the Panama island arc in Late Miocene to Recent times (Mann *et al.* 2006).

One of the best-studied faults in this zone with prominent bends is the Bocono Fault of western Venezuela, which forms one edge of the triangular Maracaibo Block that is being ejected northward along the right-lateral Bocono and the left-lateral Santa Marta Fault in Colombia to the east (Fig. 35a). The Bocono fault zone exhibits a paired bend defined by the La Gonzalez gentle releasing

bend, and an adjacent, unnamed restraining bend with high topography (Schubert 1980; Fig. 35b).

An ancient pull-apart, the Icoatea Basin, related to the early evolution of this strike-slip boundary, has been studied using 3D seismic data in the subsurface of the Maracaibo Basin in western Venezuela (Escalona & Mann 2003; Fig. 36). These studies have clearly imaged the transverse faults responsible for an Eocene pull-apart basin, which have eluded researchers in more thickly sedimented pull-aparts.

Tectonic setting and geological development of bends along the active Anatolian strike-slip fault systems

Introduction. The Anatolian region (roughly corresponding with the country of Turkey) is a well-known example of 'tectonic escape' triggered by the collision between a continental promontory on the Arabian plate and the Eurasian plate (Sengör *et al.* 1985; Hempton 1987; Fig. 37). Two major strike-slip faults, the North and East Anatolian faults, were formed about 5 Ma ago, in response to the collision, and propagated from east to west (Sarolgu *et al.* 1992; Armijo *et al.* 2002). Because the faults have opposed shear senses, they expel the Anatolian continental fragment westward into the Mediterranean Sea, where oceanic crust of the eastern Mediterranean Sea can be subducted beneath the fragment. In addition to these two major strike-slip faults, other smaller strike-slip faults were formed within the Anatolian block and exhibit fault bends of restraining and releasing types (Sengör *et al.* 1985; Kocyigit & Erol 2001).

On the basis of geological information, Sengör *et al.* (1985) inferred that large areas of Anatolia and the Caucasus were internally deformed during the collision process. Whilst this may be true prior to the formation of the large indent-linked strike-slip faults, GPS results reported by Reilinger *et al.* (2006) show that blocks adjacent to the presently active large strike-slip faults are behaving rigidly, with little evidence for the type of internal strain envisioned by Sengör *et al.* (1985). GPS measurements show that the escape path of the Anatolian Block is highly arcuate along a counterclockwise path and proceeds at very rapid rates in the range of 20–30 mm/a (McClusky *et al.* 2000; Reilinger *et al.* 2006).

One possible mechanism for pull-apart and other basin formation related to north–south constriction of the Anatolian Block is shown in Figure 8d; a more complex model involving the westward propagation of the North Anatolian Fault is given by Chorowicz *et al.* (1999). The only place along the fault where the North Anatolian fault trace is out of alignment with the plate vector is in the

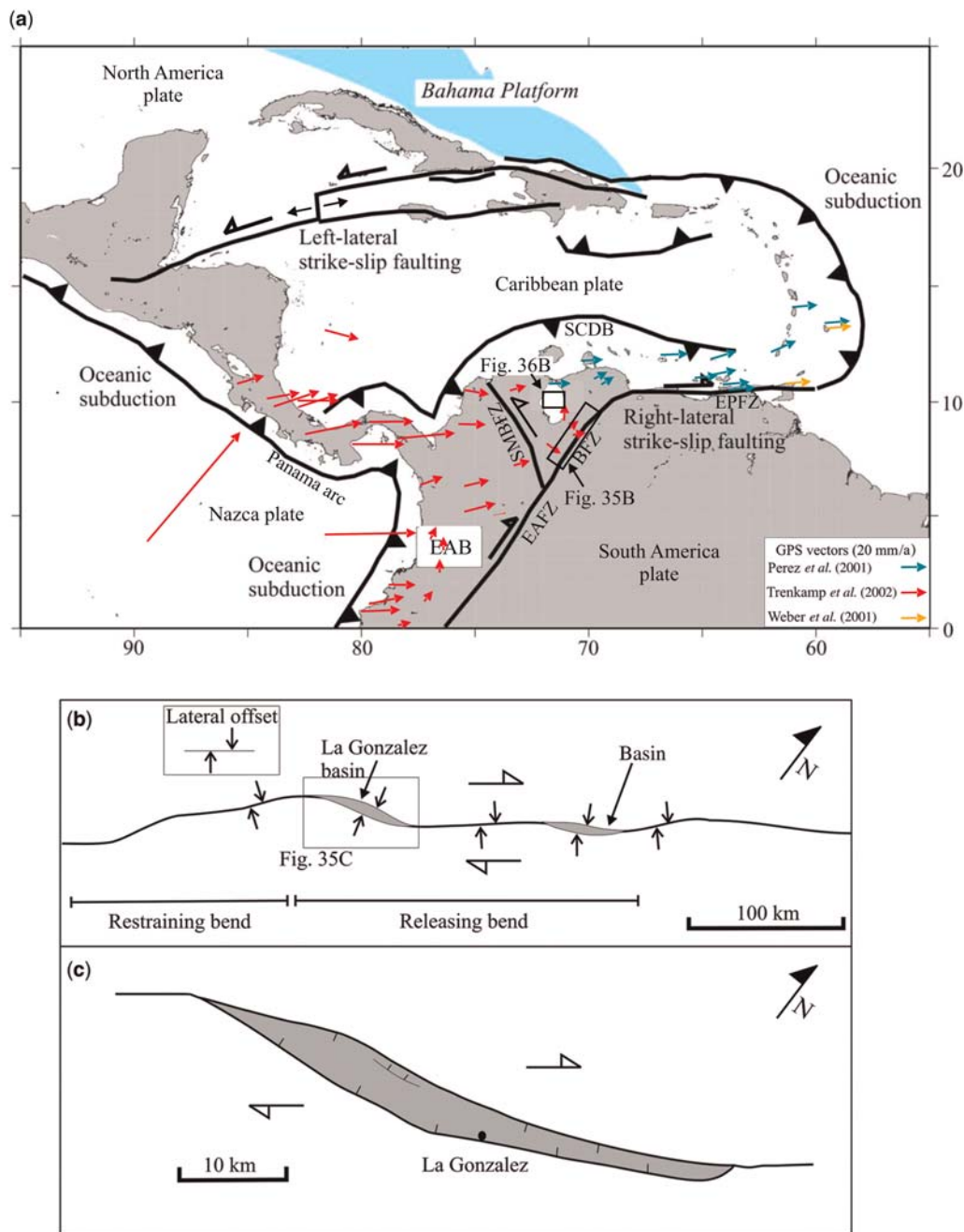


Fig. 35. (a) Tectonic map and GPS velocity field (red arrows) of northern South America, modified from Mann *et al.* (2006). Sources of the GPS data are Perez *et al.* (2001), Trenkamp *et al.* (2002), and Weber *et al.* (2001), and are colour-coded, according to author, in the inset box. The Maracaibo Block of NW South America is a large escape block activated by convergence between the Panama Arc and NW South America and moving to the north and NE along the left-lateral SMBFZ (Santa Marta–Bucaramanga fault zone) and the right-lateral Bocono fault zone (BFZ). (b) Map of the Bocono fault zone of Venezuela, modified from Schubert (1980), showing possible paired-bend structure. (c) Map of the La Gonzalez pull-apart basin similar to young pull-apart basins with a lazy-Z shape (cf. examples in Fig. 3a–f).

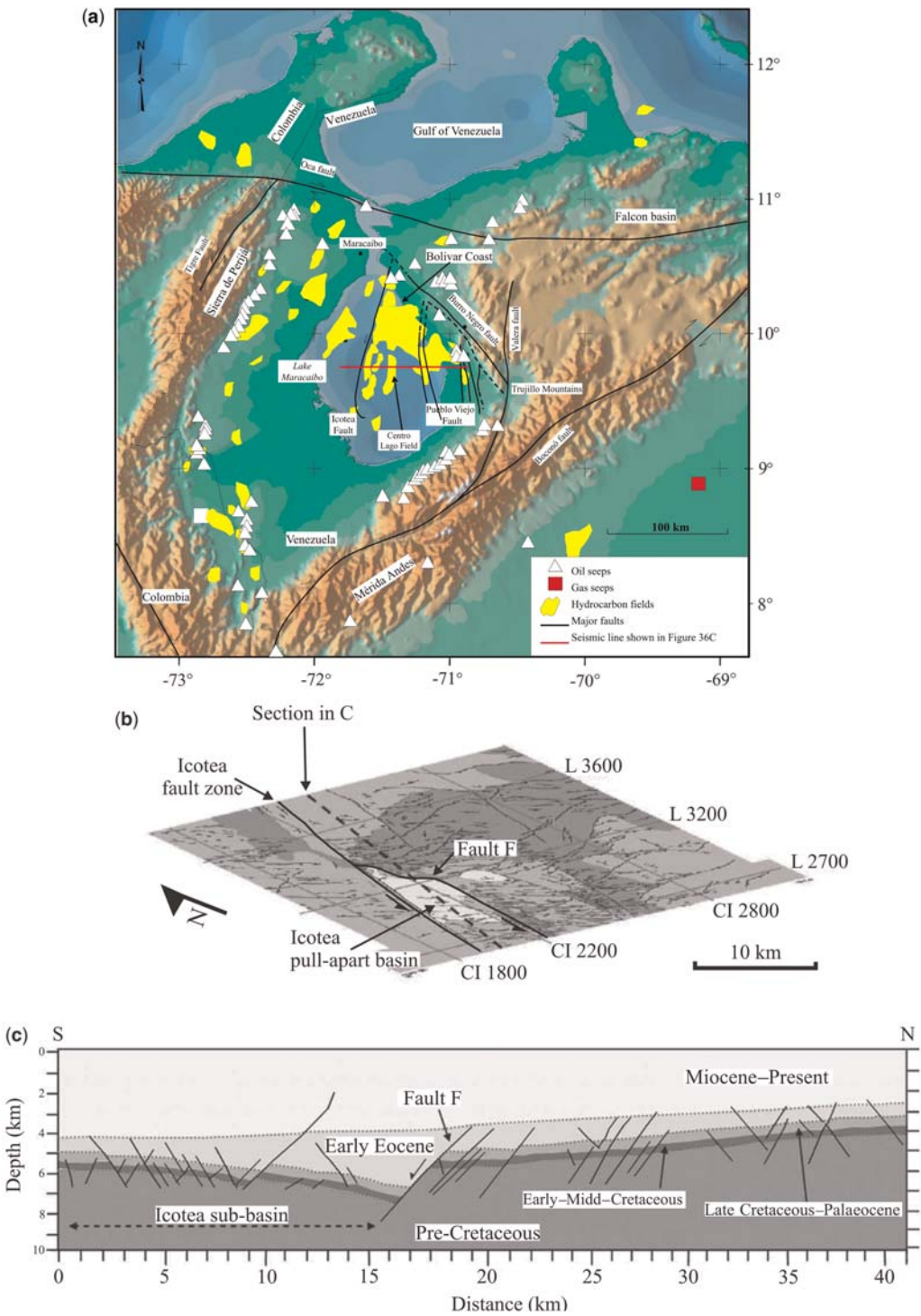


Fig. 36. (a) Tectonic setting of the Eocene Icoetea pull-apart basin in the Maracaibo basin, Venezuela (from Escalona & Mann 2003). (b) Two-way time slice from a 3D seismic survey, showing a pull-apart structure preserving younger Eocene sedimentary rocks in the left-stepping pull-apart. (c) Cross-section through the Eocene pull-apart basin, showing basin formation along normal fault F, and the inactivity of faults in post-Eocene times. Compare the structure of the pull-apart on the normal fault with that of the Ridge Basin of southern California in Figure 19.

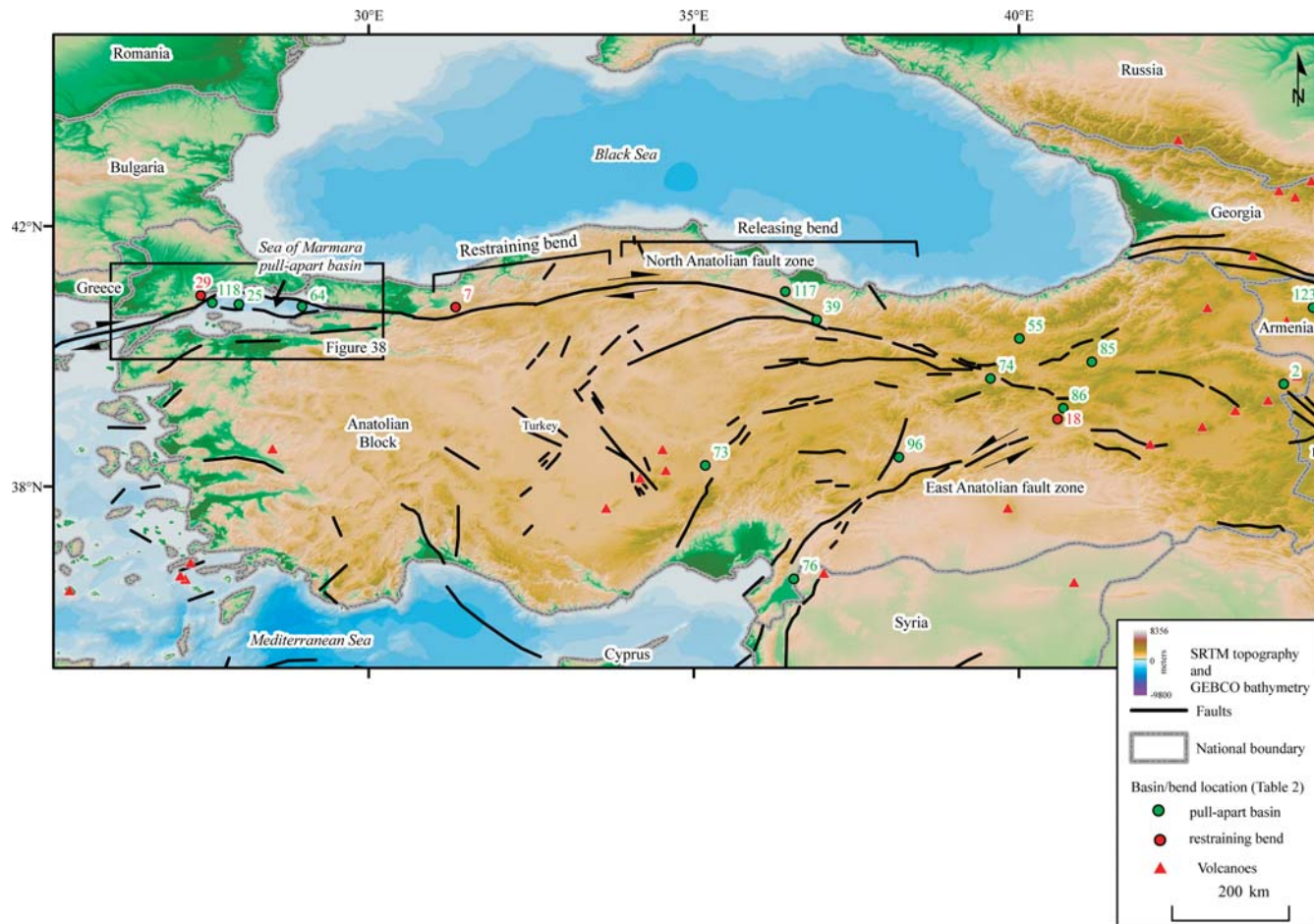


Fig. 37. (a) Regional tectonic map of the North and East Anatolian strike-slip fault systems of Turkey, using SRTM topography and GEBCO bathymetry as a base map. Black lines are active faults. GPS studies show that the direction of plate motion of the Anatolian block is roughly east–west and right-lateral for the North Anatolian Zone, and left-lateral for the East Anatolian fault zone. Laterally extensive paired bends are interpreted along both boundaries.

Sea of Marmara region (Fig. 38). In areas to the east, the fault is approximately parallel to the slip vectors constrained by GPS measurements (Reilinger *et al.* 2006).

Bends along the North Anatolian fault zone. Right-lateral offset estimates for the North Anatolian Fault vary widely, but many agree to its right-lateral offset estimate of 25–85 km over the past 5 Ma (cf. Armijo *et al.* 1999). Major bends along the trace of the fault visible at the scale of the map shown in Figure 38 include the Ganos restraining bend; the Sea of Marmara pull-apart basins; the Niksar Z-shaped pull-apart; and the Erzincan rhomboidal pull-apart (Barka & Gulen 1989; Fig. 38; Table 2). Smaller pull-apart and fault wedge basins, and their relationships to earthquake ruptures, are described by Barka & Kadinsky-Cade (1988), Barka (1992), Lettis *et al.* (2002) and Harris *et al.* (2002).

Bends in the Sea of Marmara. This is one of the most intensively surveyed and studied strike-slip bend areas in the world, because of the likely occurrence of a future strike-slip rupture on the North Anatolian fault zone near the large population centre and strategic port of Istanbul, Turkey, on the Bosphorus Strait linking the Sea of Marmara with the Black Sea (Fig. 38). The last rupture, in 1999, on a century-long series of westward-propagating ruptures, affected the Izmit area east of Istanbul (Harris *et al.* 2002; Lettis *et al.* 2002; Fig. 38). In this area, the North Anatolian fault zone bifurcates into two strands: a northern strand is marked by an array of three spindle to lazy-Z-shaped pull-apart basins in the Sea of Marmara, and a large restraining bend along the Gallipoli Peninsula and Ganos Mountain marked by en échelon folding (Armijo *et al.* 1999; Fig. 38).

There have been many surveys of the three pull-aparts in the Sea of Marmara, with most controversy focused on whether the basins are active or have been bypassed by through-going strike-slip fault traces (Okay *et al.* 2000; Rangin *et al.* 2004; Seeber *et al.* 2006). A straighter and more southerly strand of the fault called the southern strand of the North Anatolian fault zone passes south of the Sea of Marmara. GPS surveys show that motion is roughly east–west and parallel to the slight arc in the strike-slip faults (Fig. 38).

Paired-bend interpretation. The pattern of adjacent restraining and releasing bends shown on Figure 38 can be interpreted in terms of the same paired-bend model used for the Californian and Lebanese bends. Deviation of the northern strand by the restraining bend, is returned to the original trend of the fault trace by the opening of the pull-

apart array in the Sea of Marmara. The southern strand may represent a single trace of the North Anatolian fault prior to this northern deviation, since GPS vectors indicate that most of the present-day motion is occurring on the northern trace of the fault (Fig. 38).

Bends along the East Anatolian fault zone. The East Anatolian fault zone exhibits a very discontinuous fault trace, with one gentle restraining bend near Bingol and one rhomboidal pull-apart at Lake Hazar (Hempton & Dunne 1983; Table 2). Smaller pull-apart and fault wedge basins and their relationships to earthquake ruptures are described by Barka & Kadinsky-Cade (1988) and Barka (1992).

Tectonic setting and geological development of bends along the active and ancient Alpine–Carpathian strike-slip fault systems and inverted faults of the European foreland

Introduction. The Alpine–Carpathian collision consisted of a west to east, diachronous continent–continent collision from France in western Europe to Romania in central Europe (Royden *et al.* 1982; Ratschbacher *et al.* 1991; Fig. 39). This collision produced lateral escape of continental fragments towards uncollided areas in the east – along with indent-related strike-slip faults within the orogenic belt and pull-apart basins, and restraining bends at complex fault bends. In addition to deformation within the orogenic belt, far-field effects of the collision reactivated existing basement structural features of various types as far north as the British Isles and the North Sea. In this section, I review a few of the better-studied restraining and releasing bends formed as a result of this orogenic event.

Vienna Basin. Terminal collisions in western areas and continuing collisions in eastern areas caused the overriding Alps–Pannonian plate to fragment along large strike-slip fault systems (Royden *et al.* 1982; Fodor 1995; Mann *et al.* 1995; Fig. 40a). A major block boundary formed between an inactive and terminally collided area west of the Vienna area and the Pannonian plate that continued to move to the NE east of the Vienna area (Fig. 40b). As offset lengthened along this boundary, a large rhomboidal basin formed beneath the site of the present-day city of Vienna, Austria (Fig. 40c).

Surface mapping of scarps, subsurface mapping and well-log correlation by Decker *et al.* (2005) and Hinsch *et al.* (2005) has allowed a detailed picture of this rhomboidal basin with its distinctive twin depocentres characteristic of rhomboidal

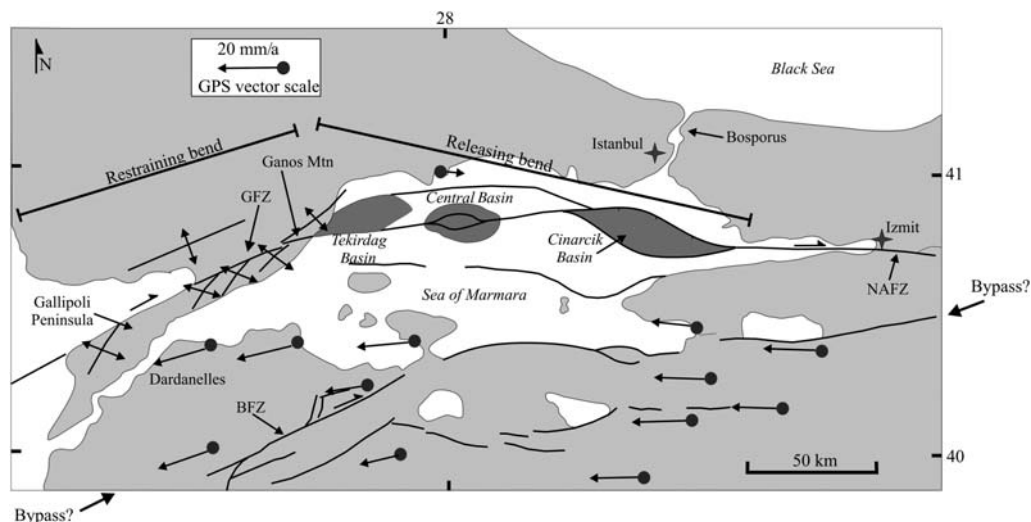


Fig. 38. Schematic map of the active traces of the North Anatolian fault system in western Turkey, modified from Armijo *et al.* (2005), and their relation to GPS plate motions, from McClusky *et al.* (2000). See the regional tectonic map in Figure 37 for location. Inferred paired restraining- and releasing bends are indicated on both subparallel fault systems. Note how the coastlines of the Sea of Marmara mirror the coastal and offshore paired-bend structure.

pull-aparts (Mann *et al.* 1983; Fig. 7d). Other pull-aparts have been described to the south (Sachsenhofer *et al.* 2003) and residual seismic activity and surface faulting continues on all of these structures to the present day.

Bends in the North Sea and British Isles. Many pre-existing structural grains were reactivated during the Early Cenozoic as a result of the Alpine collision. Some of these examples have been particularly well imaged by subsurface seismic and well data collected by the oil industry. For example, the Pijnacker oil field offshore of the Netherlands in the North Sea originated as an Early Cretaceous, right-lateral rhomboidal pull-apart that was subsequently inverted as a lazy-S gentle restraining bend when movement on the strike-slip fault reversed and became left-lateral (Fig. 41).

Dooley *et al.* (1999) show a rhomboidal pull-apart mapped from 3D seismic data in the subsurface of the Netherlands (Fig. 42a). The interior of the basin contains transverse normal faults that accommodate basin extension as predicted in the standard pull-apart model (Fig. 7). The master faults bounding the basin also exhibit a change in major throw as the basin depocentre is crossed (Fig. 42a).

Another example from Dooley *et al.* (1999) shows that a half-graben formed during pre-Jurassic and Jurassic rifting was modified by post-Early Cretaceous right-lateral motion on the Great Glen fault zone (Fig. 42b). Reactivation probably related to the Tertiary Alpine reactivation event produced

deepening of the basin along steeper master fault strands.

According to Holloway & Chadwick (1986), the rhomboidal Petrockstow and Bovey basins formed by about 6 km of left-lateral offset during the Early Tertiary, possibly as a result of extension in the offshore area of southern England. In Mid-Tertiary times, slip on the fault reversed to right-lateral and produced reverse faulting at the margins of the Petrockstow and Bovey basins (Fig. 43c). The likely mechanism of the Mid-Tertiary event was a far-field effect of the Alpine shortening event.

Tectonic setting and geological development of active bends of the Iranian Plateau

Active strike-slip faults and fault bends are common in central and eastern Iran (Fig. 44), and include the site of the classic mapping study by Tchalenko & Ambraseys (1970) of shear structures formed along the 25-km-long rupture zone of the 1968 Dasht-e-Bayaz earthquake (Fig. 45). Active deformation in Iran is related to about 25 mm/a of active convergence between the Arabian plate and Eurasia; the Arabia plate acts as a rigid indenter and produces cross-cutting strike-slip faults that separate eastern Iran from Afghanistan (cf. the Neyband, Gowk, Sistan fault zones of Walker & Jackson 2004; Fig. 44). Some shortening is absorbed by folding in the Zagros fold-and-thrust belt, whilst the remainder is absorbed by conjugate

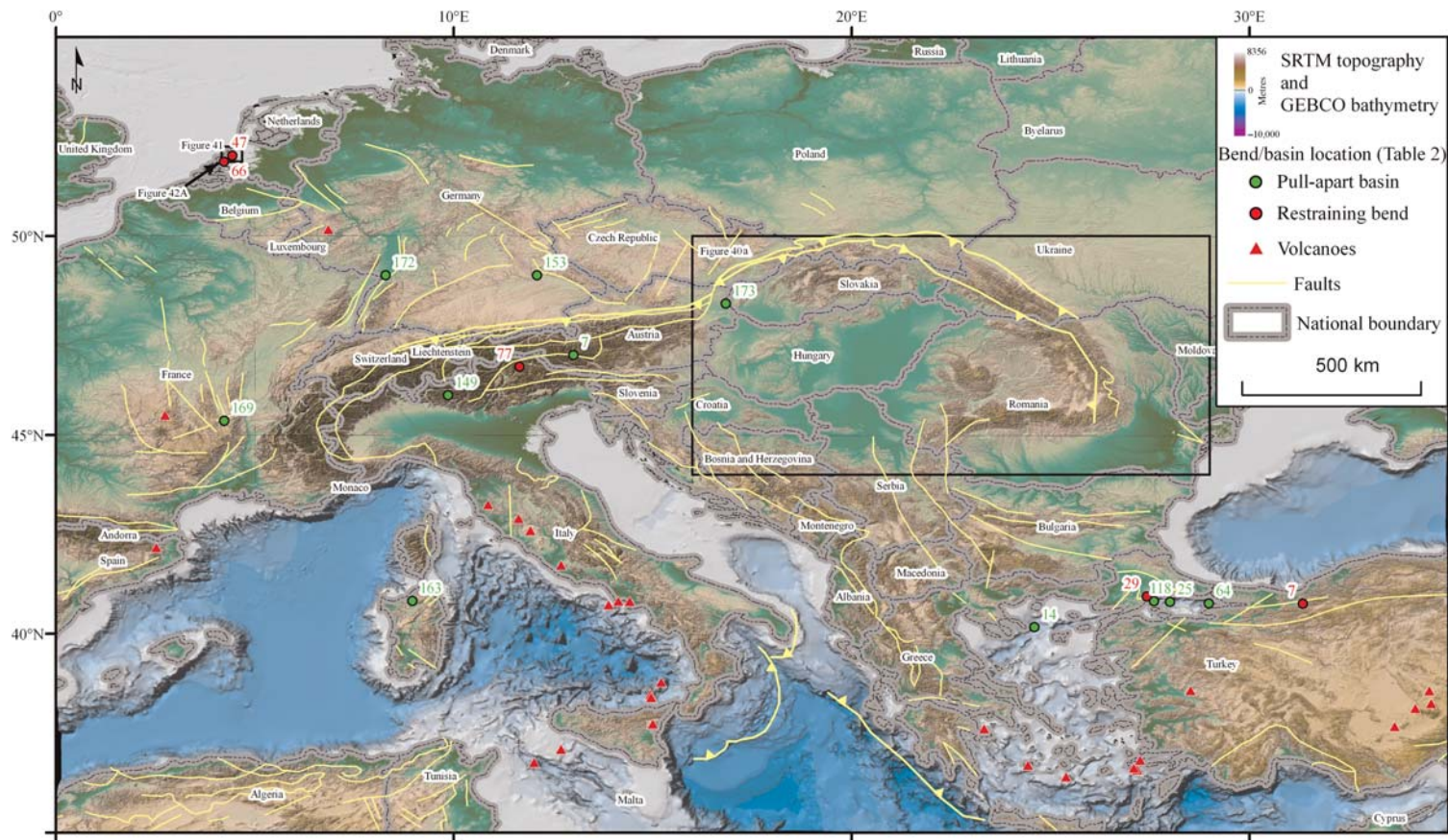


Fig. 39. Regional tectonic map of Cenozoic to recent strike-slip fault systems of Europe, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilations by Levine (1995) and Coffin *et al.* (1997).

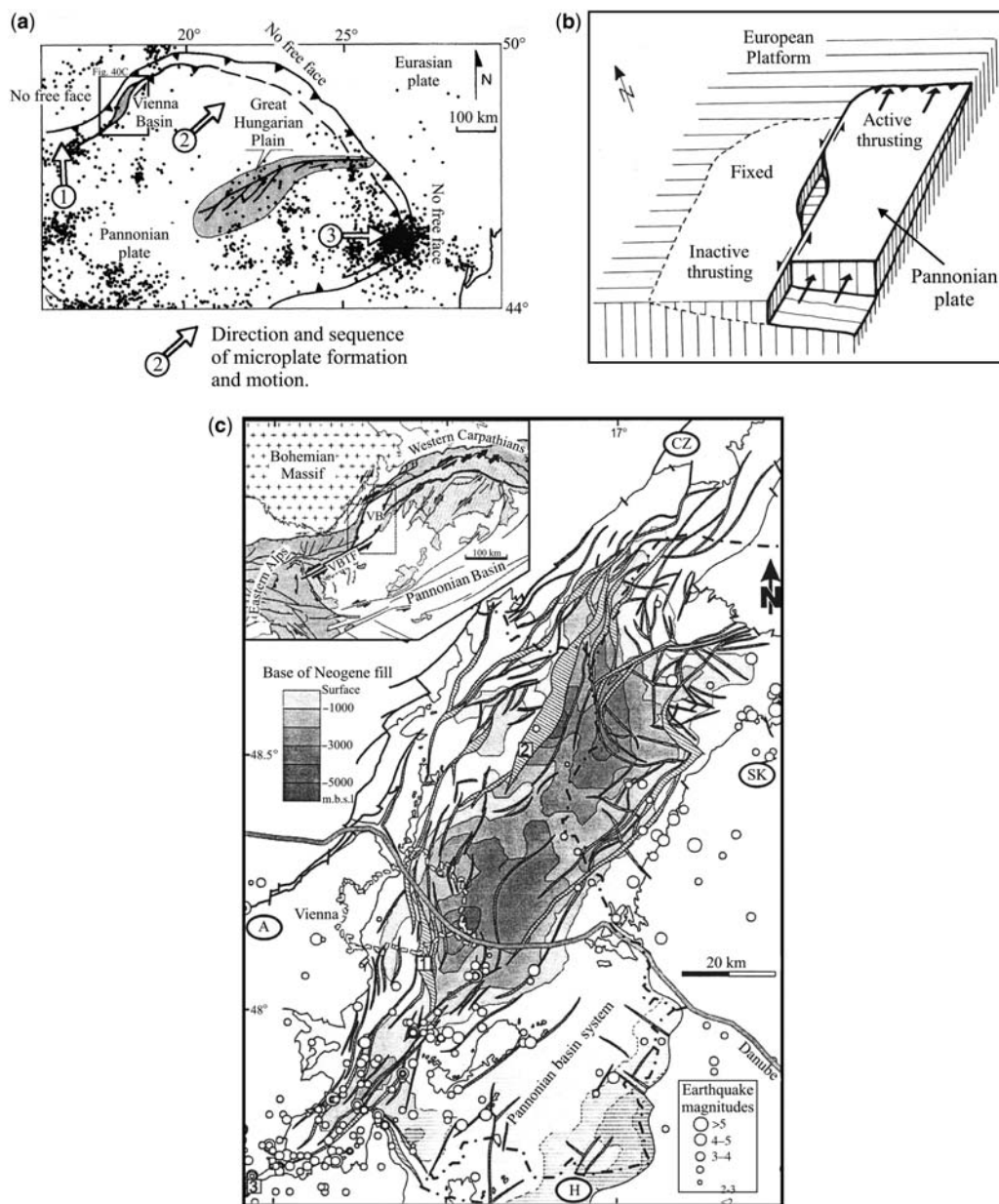


Fig. 40. (a) Tectonic setting of the Vienna Basin during west-east oblique collision, from Mann *et al.* (1995). (b) Block diagram, showing relation between the Vienna pull-apart basin and underlying nappes and thrust faults (from Royden 1985). (c) Structural setting and seismicity of the Vienna pull-apart basin, from Hinsch *et al.* (2005). Inset shows main structural units of the Eastern Alpine-Carpathian region. Key Abbreviations: VB, Vienna Basin; VBTF, Vienna Basin transfer fault. The map shows the faulted pre-Neogene basement surface and earthquake epicentres since 1201 BC. Labelled points: 1, Leopoldsdorf Fault; 2, Steinberg Fault; 3, transition to the Mur-Murztal Fault. A, Austria; CZ, Czech Republic; H, Hungary; SK, Slovakia.

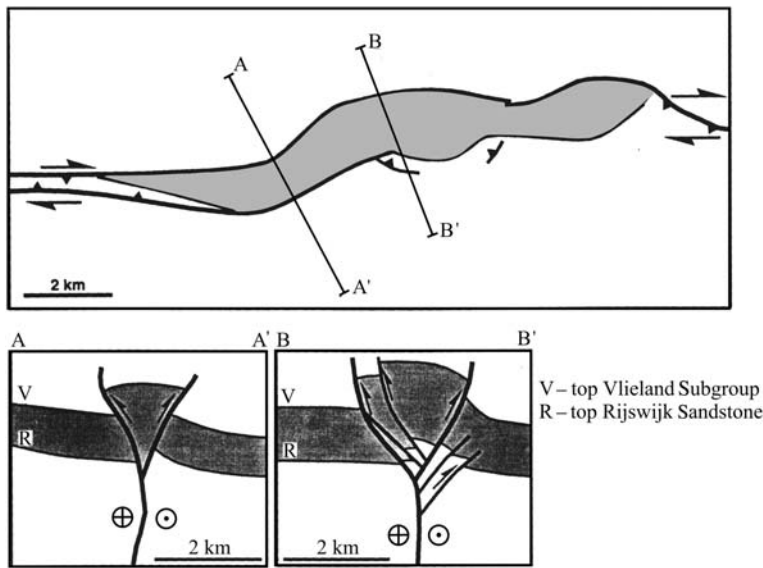


Fig. 41. (a) Structure map of the Pijnacker field in the west Netherlands, using subsurface mapping from Racero-Baema & Drake (1996) (see the location of the map on Fig. 39). This structure formed as an Early Cretaceous rhomboidal pull-apart basin, and was inverted during the Early Tertiary. (b) Cross-sections based on seismic-reflection profiles.

strike-slip faults and block rotations (Walker & Jackson 2004). The left-lateral Dasht-e-Bayaz Fault is antithetic to the overall right-lateral shear zone and accommodates a component of clockwise block rotation within the overall right-lateral shear zone (Fig. 45).

Whilst a paired bend is present at the scale of the Dasht-e-Bayaz rupture, few pull-aparts are observed at more regional scales in this area. Restraining bends commonly connect different strands within the various shear zones, such as the lazy-S gentle bend along the Purang right-lateral strike-slip fault in eastern Iran (drawn from Fig. 7 of Walker & Jackson 2004) (Fig. 46a). The $M_{6.6}$ 2003 Bam earthquake, which caused over 40 000 deaths, originated on a buried right-lateral fault parallel to the Gowk fault zone in eastern Iran. Imagery studies show that the rupture occurred at the edge of a gentle restraining bend (Fu *et al.* 2004).

Tectonic setting and geological development of active bends of the Tibetan Plateau, western China and Mongolian Altai

Continental collision effects are most prominent in central and eastern Asia, where the India indenter is forcing tectonic blocks to the east and SE along indent-linked strike-slip faults (Fig. 47) (Tapponier *et al.* 1982). The northern and eastern boundaries of

the escaping continental fragments are mainly left-lateral faults, whereas the southern and western boundaries are mainly right-lateral faults (Fig. 47a). About half of the 50 mm/a of northward convergence between India and Asia is absorbed by the horizontal extrusion of the escaping fragments (Avouac & Tapponnier 1993). The area affected by collision extends into western China, Mongolia, Thailand, the Andaman Sea, and perhaps as far north as Siberia.

Figure 47a shows a kinematic model for the deformation of Asia, based on various types of geological, earthquake, and tectonic data compiled by Avouac & Tapponnier (1993) and redrawn by Yeats *et al.* (1997). The rotation senses of major blocks are given along the major directions of block motions. An obvious characteristic of the indent-linked strike-slip faults is that they are generally curved and discontinuous and therefore contrast with the more regular continental-boundary faults discussed earlier. Another characteristic is the absence of large-scale pull-apart zones or paired bends.

Fault bends of the Tibetan Plateau and western China. The largest fault on the Tibetan Plateau is the 600-km-long left-lateral Altyn Tagh Fault, that is believed to have initiated in the early Eocene (Fig. 47b). Cowgill *et al.* (2004a, b) studied the isolated Akato Tagh gentle bend occurring at a

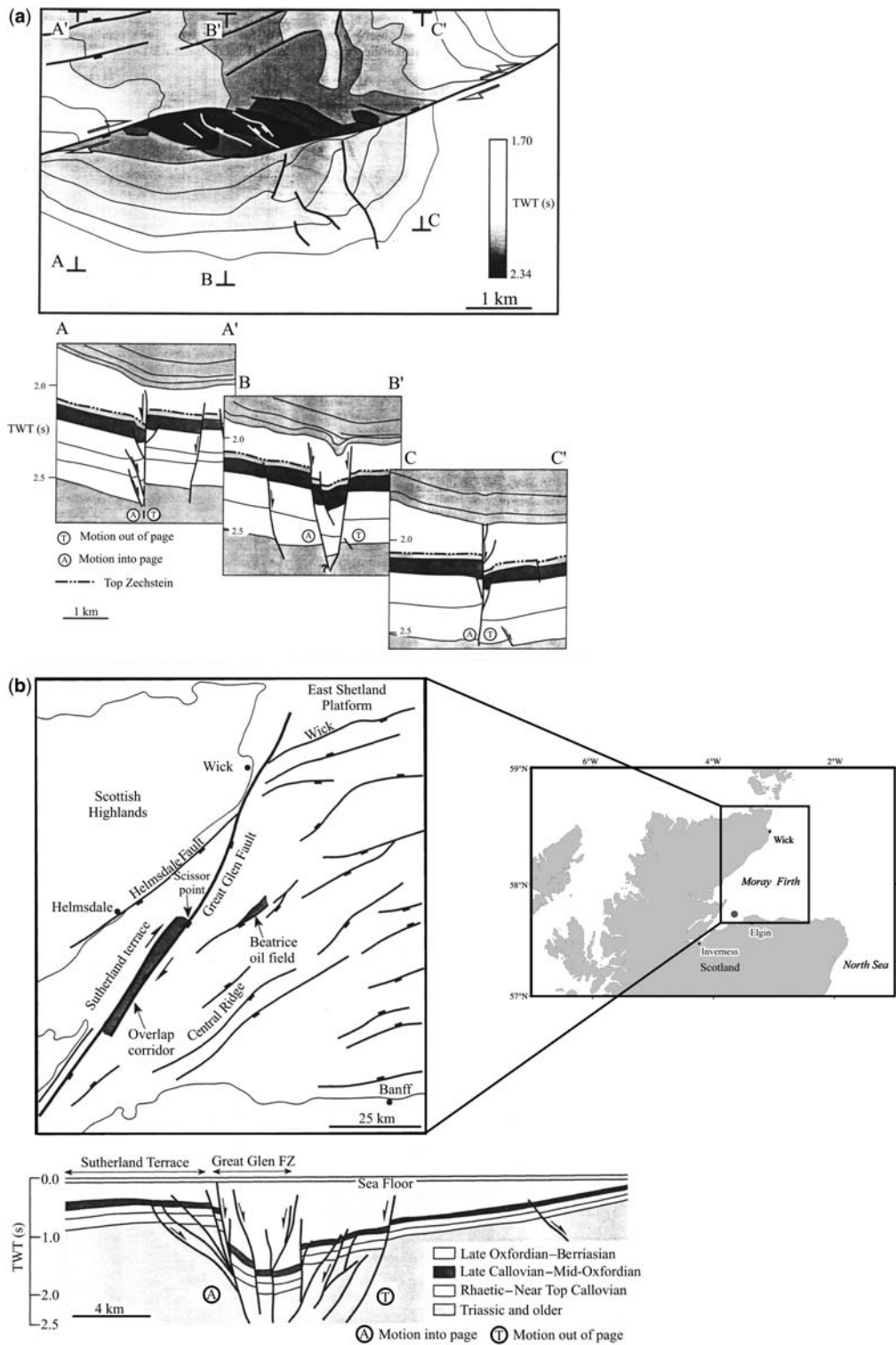


Fig. 42. (a) Time-structure map and cross-sections from a rhomboidal basin in the Netherlands, using 3D seismic-reflection data from Dooley *et al.* (1999) and subsurface mapping from Richard *et al.* (1995). (See location of the map on Fig. 39.) (b) Map and cross-section of a rhomboidal pull-apart basin of Late Jurassic age in the Inner Moray Firth, from Dooley *et al.* (1999), using subsurface mapping of Underhill (1991).

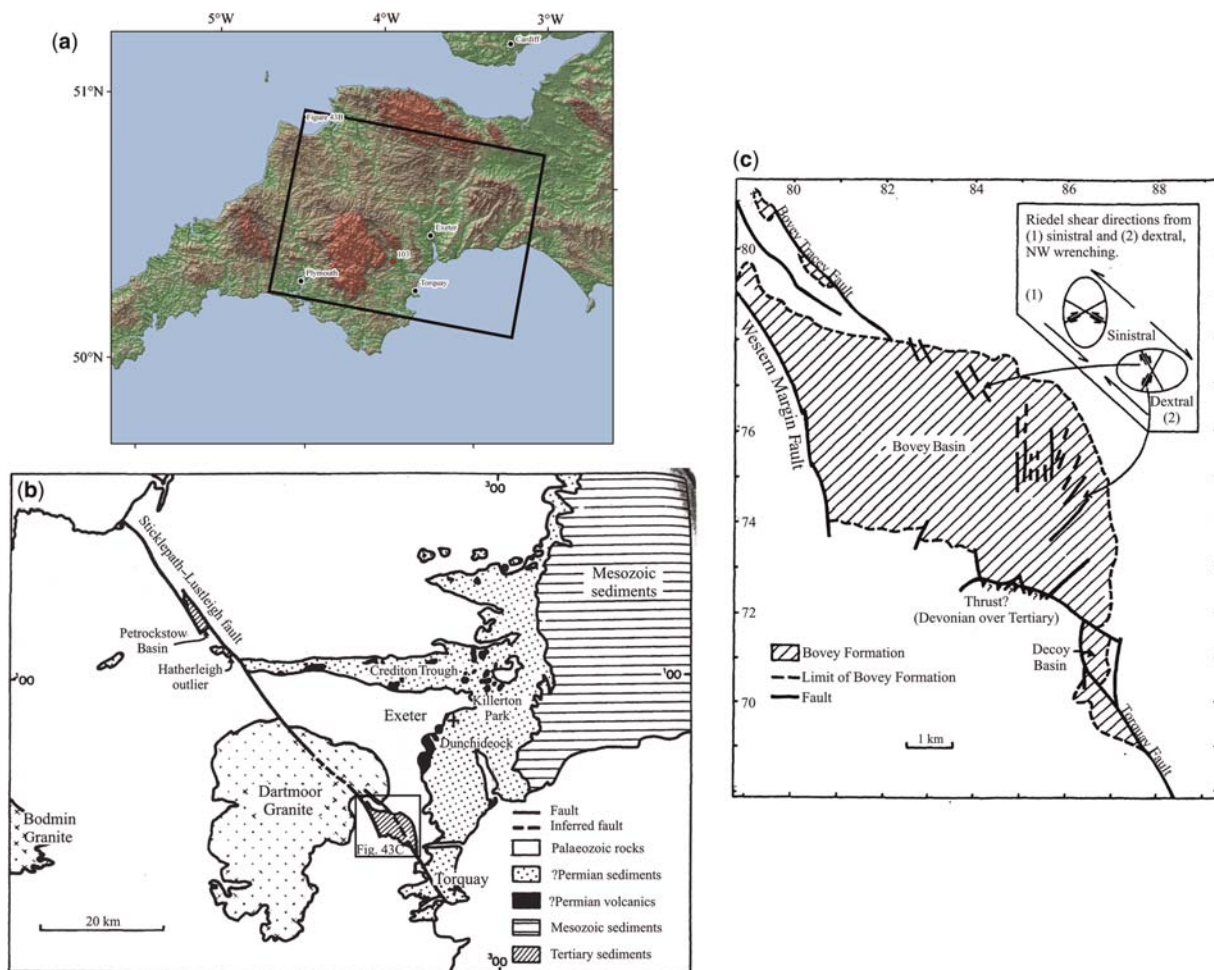


Fig. 43. (a) SRTM topography and GEBCO bathymetry used as a base map. (b) Simplified geological map of part of SW England, showing the Sticklepath–Lustleigh fault system, modified from Holloway & Chadwick (1986). (c) Geological map of the Bovey Basin, the largest basin along the Sticklepath–Lustleigh fault system. The basin formed by left-lateral offset on the left-stepping Bovey Tracey–Torquay faults and the Western margin fault. Minor Late Tertiary reactivation of the Bovey Tracey–Torquay faults is indicated by the presence of minor faults that have been interpreted as Riedel Shears.

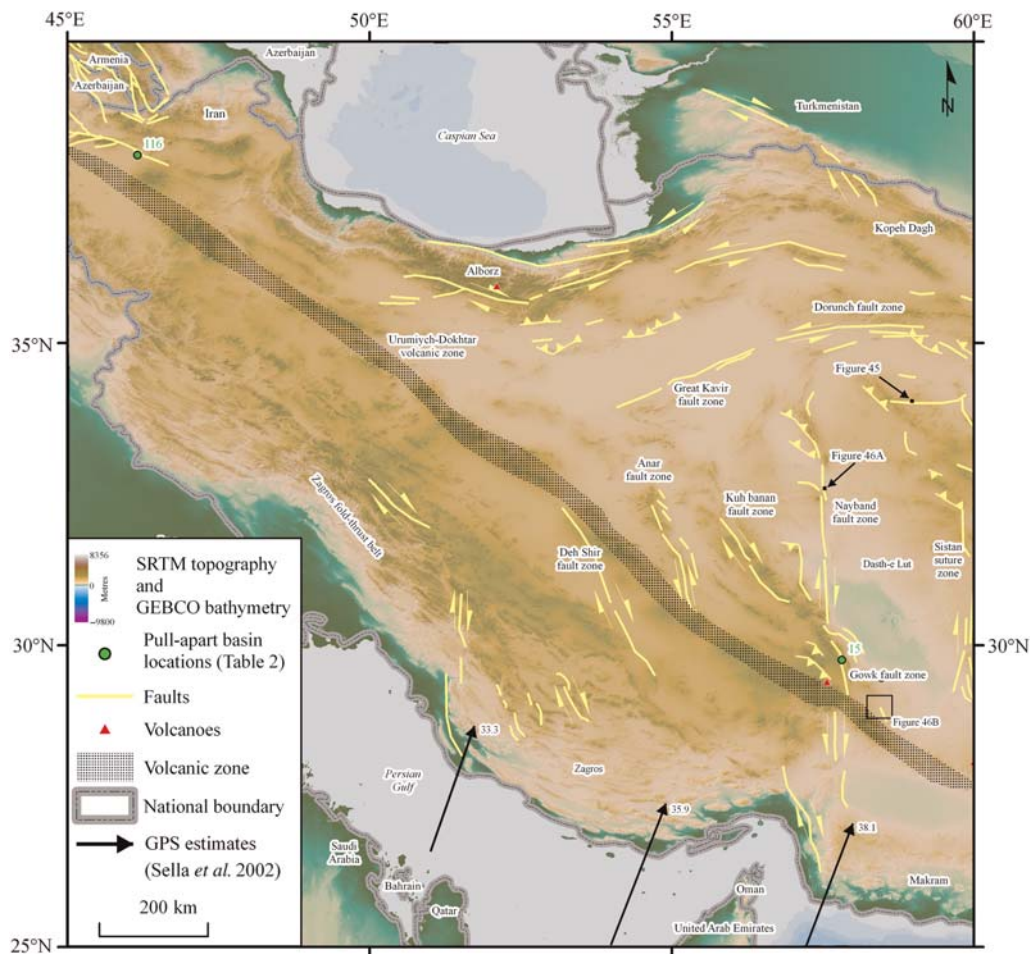


Fig. 44. Regional tectonic map of the active to Cenozoic strike-slip fault systems of Iran, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults, taken from compilations by Walker & Jackson (2004) and Karakhanian *et al.* (2004). GPS-based directions of Arabia–Eurasia plate motion are from Sella *et al.* (2002).

20-km-wide fault separation (Fig. 48). Their main conclusions were that the bend and its zone of transpressional deformation probably formed recently, as the result of linkage of an en échelon set of shear fractures during the recent development of this fault strand.

Pull-apart basins are more common on right-lateral faults with NE trends, like the Karakoram fault zone, where an array of two large lazy-Z shape basins are present (Murphy & Burgess 2006); the Gulu fault zone, where one rhomboidal pull-apart is present (Edwards & Ratschbacher 2005); and the Haiyuan fault zone where an array of five rhomboidal pull-aparts are present (Zhang *et al.* 1989; Figs 47b, 49).

Mongolian and Gobi Altai. Restraining bends of various structural styles are common in the Mongolian and Gobi Altai and are described in detail by Cunningham (2003a,b; 2007). Four of these bends are illustrated on Figure 50, and they exhibit some characteristics in common with bends on the indent-linked strike-slip faults of the Tibetan Plateau. These bends are found in mechanically weak Palaeozoic accretionary belts that separate older cratonic areas and are subject to far-field stresses induced by the collision of India with Eurasia.

Restraining bends in this setting have distinctive characteristics. First, they tend to be isolated structures; second, the master faults cannot be traced far

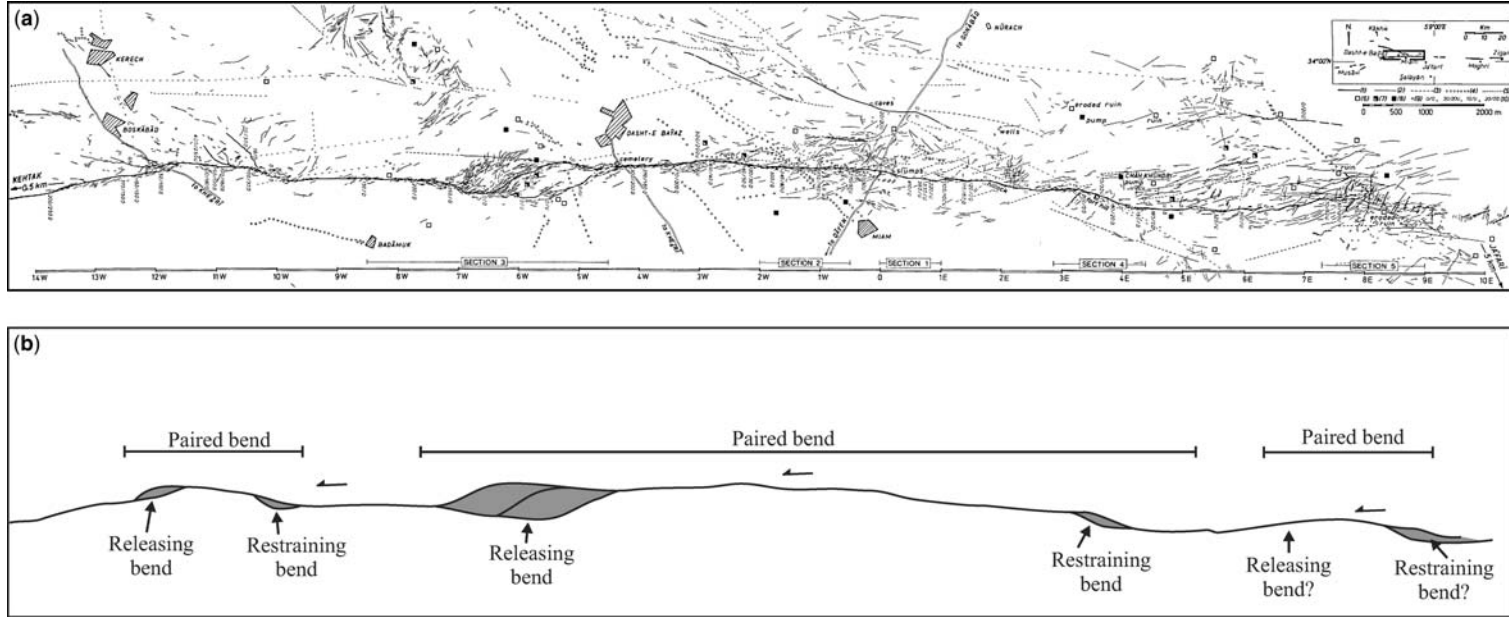


Fig. 45. (a) Fracture pattern in Quaternary alluvium that was mapped by Tchalenko & Ambraseys (1970), following the $M7.2$ Dasht-e-Bayaz earthquake of 31 August 1968 in Iran (location shown on Fig. 44). (b) Interpretation, showing the paired-bend structure.

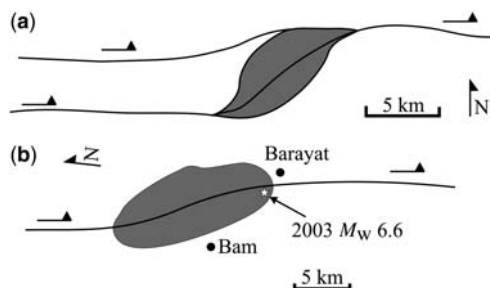


Fig. 46. (a) Sketch of the restraining bend connecting parallel strike-slip fault zones, taken from a Landsat image presented by Walker & Jackson (2004) for the Purang right-lateral strike-slip fault of Iran. (b) Sketch from the imagery of Fu *et al.* (2004) of a restraining bend near the epicentre of the 2003 M_w 6.6 Bam earthquake (star) that killed 40 000 people and left 30 000 injured.

as continuous strike-slip faults; and, third, pull-apart basins and paired bends are not recognized.

Tectonic setting and geological development of active bends of SE Asia

Several strike-slip fault systems extend SE through Myanmar, Thailand and Viet Nam, and may accommodate the escape of tectonic wedges into this region (cf. Morley 2002). Morley (2007) and Smith *et al.* (2007) have described a 150-km-long, lazy-Z restraining bend (the Chainat Complex) on the 500-km-long left-lateral Mae Ping fault zone.

Tectonic setting and geological development of active bends of the Lake Baikal region, north-central Asia

Oligocene to Recent rifting and basin formation in Lake Baikal (Fig. 51) can be explained by a step-over in a roughly east–west striking left-lateral fault zone that closely follows the southern edge of the Siberian craton. Extension in this region was long considered to be a distant, indentation and escape effect of the India collision, but Calais *et al.* (2003) have recently challenged this idea, using GPS results from the region. Their view is that trench forces from the western Pacific may dominate over the distant indent-linked India event.

Lake Baikal Basin. This seismically active basin is a lazy-S-shaped pull-apart basin occupied by a deep lake. The basin formed in the Oligocene and remains an active feature up to the present day (Mats 1993; Petit 1998; Calais *et al.* 2003).

Tunka Basin. The Oligocene to Recent Tunka Basin forms one of the many satellite basins surrounding Lake Baikal, and is an excellent example of a lazy-Z-shaped pull-apart basin (Larroque *et al.* 2001; Arjannikova 2004; Fig. 52).

Tectonic setting and geological development of ancient bends of the mid-continent of North America

Strike-slip faulting was pervasive in the interior of North America during the Late Palaeozoic collision between northern South America and the southern edge of North America and during the Late Cretaceous–Early Tertiary Laramide event associated with shallow subduction or collision along the western edge of North America (Kluth & Coney 1981; Marshak *et al.* 2003; Fig. 53a–c). A modern analogue of this type of distributed intra-continental deformation would be the modern India–Asia collision (Fig. 53c).

Owl Creek and Quealy Dome restraining bend. Bends in this area of Wyoming appear to be reactivated by the NE-directed Laramide event (Fig. 53d). The Owl Creek uplift is a restraining bend along an east–west strike-slip fault (Paylor & Yin 1993; Fig. 54a & b). The Quealy Dome formed between two left-lateral strike-slip faults related to the Laramide Orogeny (Fig. 54c & d) and resembles bends created in laboratory models (McClay & Bonora 2001).

Arbuckle Mountains and Ardmore Basin. Deformation and restraining-bend formation in this area was related to the inversion of the Cambrian Southern Oklahoma aulacogen during the Late Palaeozoic Ancestral Rockies Orogeny (Marshak *et al.* 2003; Figs 53 & 55).

Tectonic setting and description of natural bends on strike-slip faults in zones of active and ancient oblique subduction

In his pioneer study, Fitch (1972) first pointed out the role of ‘trench-linked’ (Woodcock 1986) or ‘trench-parallel’ (Yeats *et al.* 1997) strike-slip faults in accommodating the strike-slip component of oblique subduction zones. The basic principle is that a vertical strike-slip fault can concentrate shear in a much more efficient way than distributing shear across a much larger subhorizontal subduction zone interface. Fitch (1972) also noted that trench-linked faults probably nucleate on weak

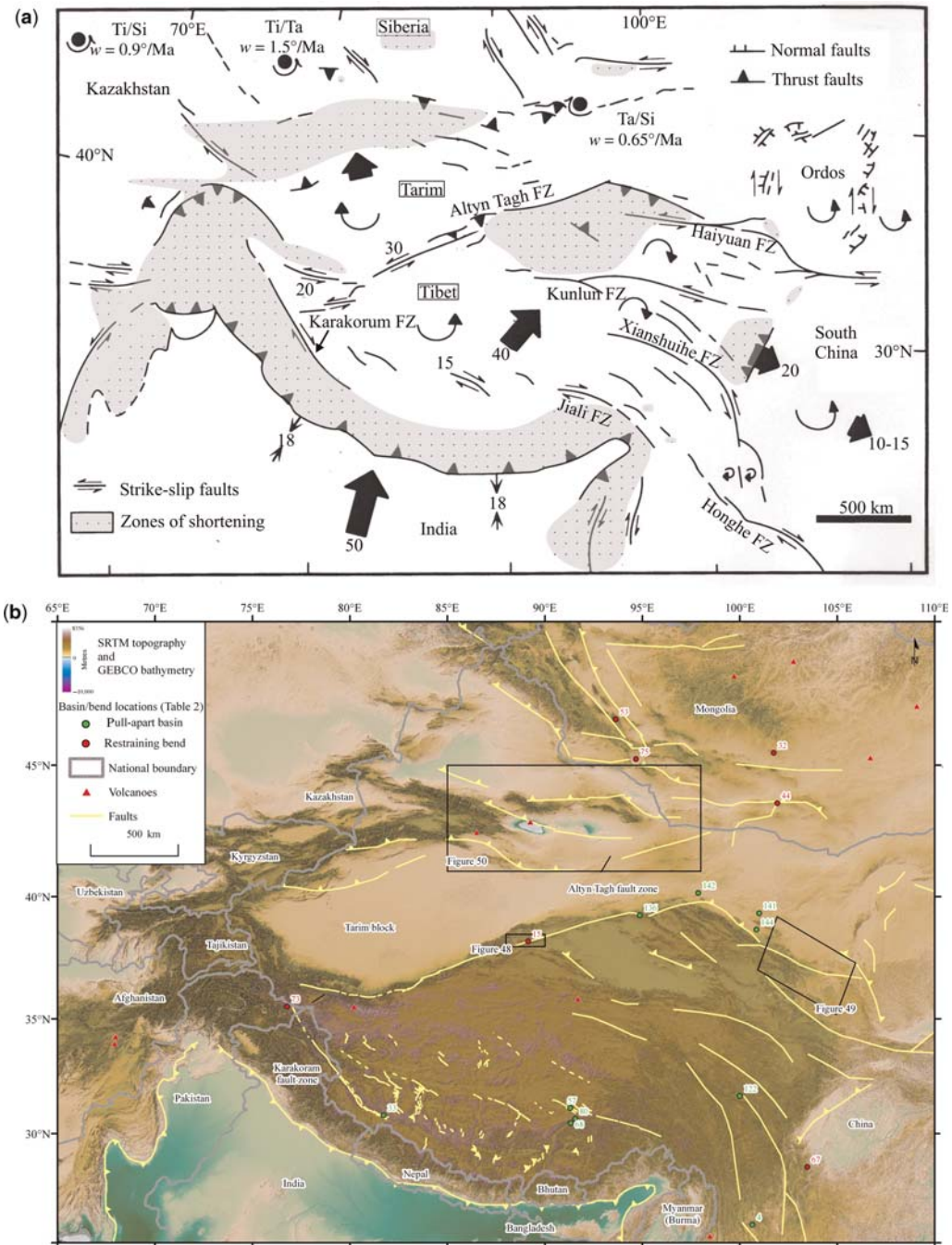


Fig. 47. (a) Block rotation model for the deformation of Tibet, from Yeats *et al.* (1997), redrawn from Avouac & Tapponnier (1993). Large arrows show block directions, small arrows show directions of rotation, and grey stippled zones show areas of shortening. (b) Regional tectonic map of active to Cenozoic strike-slip fault systems of the Tibetan Plateau and Central Asia, using SRTM topography as a base map. Yellow lines are active faults taken from compilation by Howard *et al.* (2003) and Levine (1995). Boxes show locations of the detailed maps in Figures 48–50.

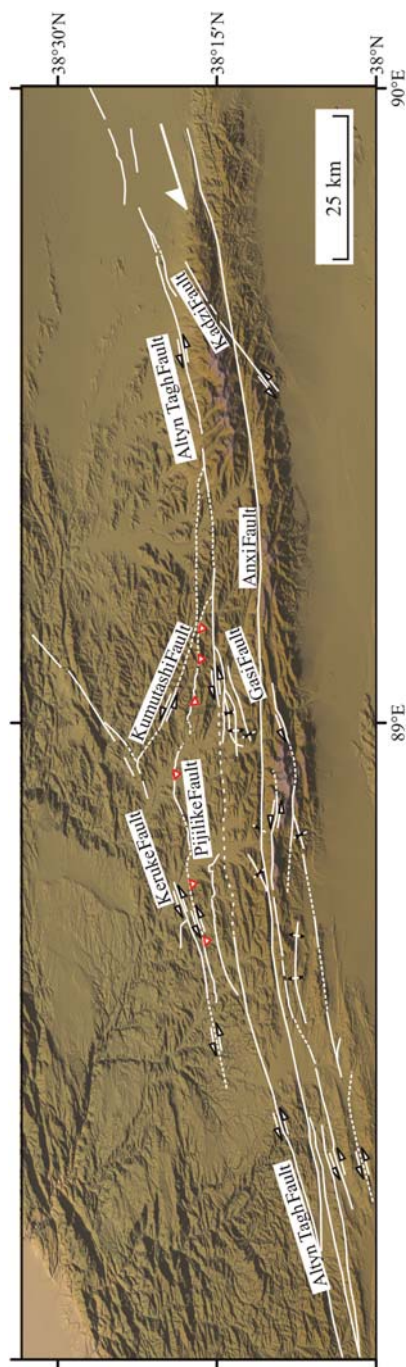


Fig. 48. Major faults of the Akato Tagh restraining bend along the Altyn Tagh fault zone, shown on an STRM base map and modified from Cowgill *et al.* (2004a, b). See Figure 47 for location.

zones in the crust formed by heated and thinned crust of the volcanic arc itself.

Jarrard (1986) summarized all aspects of trench-parallel strike-slip faults, and introduced the term 'fore-arc sliver plate' to refer to the micro-plate defined by the trench axis and the trench-parallel strike-slip fault. His review of trench-linked strike-slip faults has been updated with more recent data by Yeats *et al.* (1997). In this section, I will present some examples of trench-linked strike-slip faults with restraining- and releasing-bend segments, in order to compare these bends with those along continental-boundary and indent-linked strike-slip systems. In general, trench-linked faults are highly dependent on the trend of the subduction margin, and will change their character and rate as a function of changes in the strike of the subduction zone. For that reason, trench-linked faults tend to be less continuous and more segmented than continental-boundary faults.

Tectonic setting and geological development of active bends of the Sumatran strike-slip fault system, Indonesia

The right-lateral Sumatran fault system extends 1900 km as a sinusoidal and segmented fault zone (Sieh & Natawidjaja 2000) from its terminus as a Z-shaped pull-apart basin in the Sunda Strait between the islands of Sumatra and Java (Pramumijoyo & Sebrier 1991; Leglemann *et al.* 2000) to an extreme rhomboidal pull-apart basin floored by oceanic crust in the Andaman Sea (Curry 2005; Fig. 56). The rate of right-lateral fault slip passes from zero at its terminus in the Sunda Strait to a rate of 60 mm/a in the Andaman Sea.

A 300-km-wide fore-arc sliver plate occupies the space between the Sumatran trench and deformation front and the Sumatran Fault (Fig. 56). The overall shape of the fault is sinusoidal, with the northern half of the fault concave to the SW and the southern half of the fault concave to the NE. Sieh & Natawidjaja (2000) note that the sinusoidal trace of the Sumatran Fault is mimicked by the sinusoidal trace of the Sumatran deformation front, indicating a close genetic link between the subduction process and the geometry of the trench-linked fault. The question is whether deformation of the subducted slab influences the fault geometry in the overriding plate, or whether the fault geometry of the overriding plate has influenced the subducting slab.

Near its southern end, the fault curves towards the offshore trench and the fore-arc sliver plate thins in map view. Sieh & Natawidjaja (2000) used these relationships to calculate about 100 km of stretching of the fore-arc sliver plate parallel to

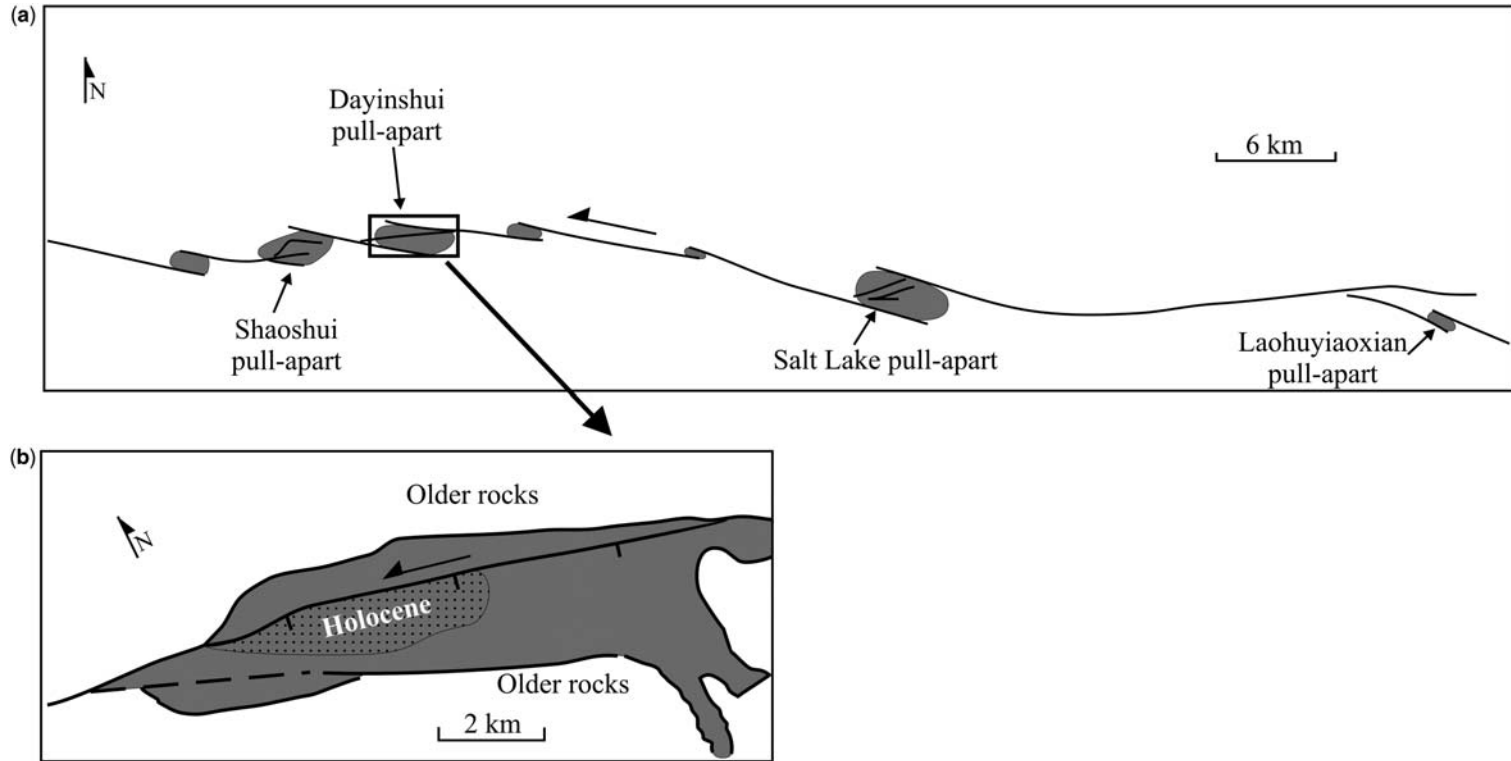


Fig. 49. (a) Map of an array of seven pull-apart basins along the Haiyuan Fault, mapped by Zhang *et al.* (1989). See Figure 47 for location. (b) The Dayinshui Basin shows 'pull-apart extinction', or an active fault cutting obliquely across the basin as observed in other basins.

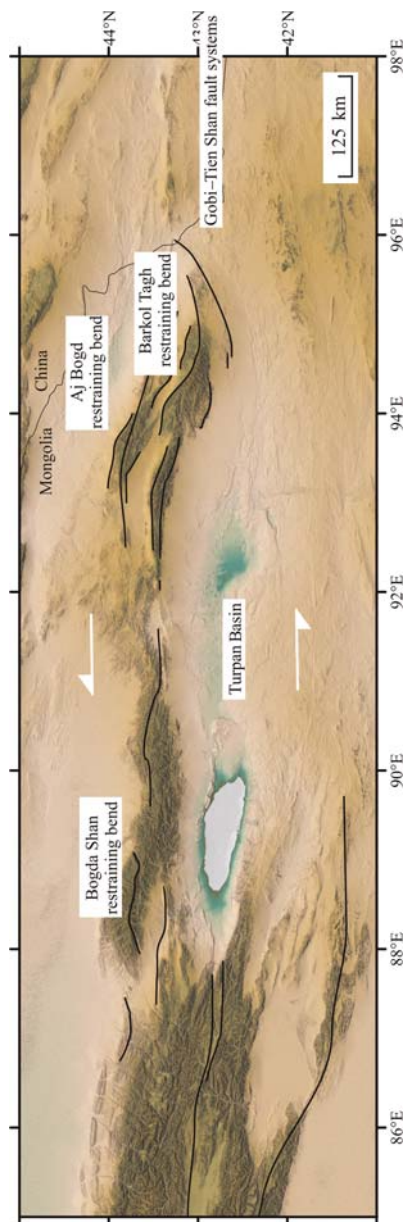


Fig. 50. Restraining bends along the Gobi-Tien Shan fault system in the Gobi Altai, southern Mongolia, modified from Cowgill *et al.* (2004a, b). See Figure 48 for location.

the trend of the Sumatran Fault. More orthogonal subduction in the area south of the Sunda Strait along the Java Trench does not require the formation of a trench-linked strike-slip fault along that segment of the trench.

Bends. A paired bend is present in the central segment of the Sumatran Fault, and is defined by a large lazy-S restraining bend with an adjacent array of four pull-apart basins showing a range of spindle-shaped to rhomboidal morphologies (Fig. 57a). This bend overlies the approximate down-dip projection of the Investigator fracture zone, a positive bathymetric feature on the down-going plate (Fig. 56). A possible bypass fault is present near the most curved part of the restraining bend (Fig. 57a).

The second-largest and best-studied of the pull-apart basins in the paired bend is the Singkarak pull-apart, which is occupied by a large lake (Fig. 57b). The 23-km length of the basin is shown by Sieh & Natawidjaja (2000) to have formed as a typical rhomboidal pull-apart (Fig. 7). Its length of 23 km is taken by them to be the approximate magnitude for the most recent phase of offset on the Sumatran Fault.

Tectonic setting and geological development of active bends of Atacama strike-slip fault system, Chile

As reviewed by Dewey & Lamb (1992) fore-arc tectonics and trench-linked strike-slip faulting are highly dependent on the orientation of the subduction zone and the obliquity of subduction along the Andean margin. The Atacama fault zone of northern Chile forms a linear zone of Quaternary left-lateral strike-slip faulting that stretches for a distance of 1100 km at an average distance of 30–50 km from the coastline (Fig. 58). A parallel, 700-km-long fault, the Coastal Scarp, exhibits evidence for Quaternary normal faulting. The Atacama Fault, has a long history of left-lateral strike-slip displacement that has been dated back to the Jurassic and Early Cretaceous (Scheuber & Andriessen 1990). Mesozoic extensional duplexes in deeper crustal rocks have been described along the Atacama fault zone by Cembrano *et al.* (2005), along with strike-slip-related block rotations (Arrigada 2003).

Armijo & Thiele (1990) point out that the observed evidence for Quaternary left-lateral shear and normal faulting contradicts the plate-tectonic prediction of right-lateral shear for oblique subduction along this part of the subduction zone (Fig. 58). They propose an alternative block model related to left-lateral shearing of blocks northward into the Arica bend (Atacama fault zone) and normal faulting above the underlying thrust stacks (Coastal

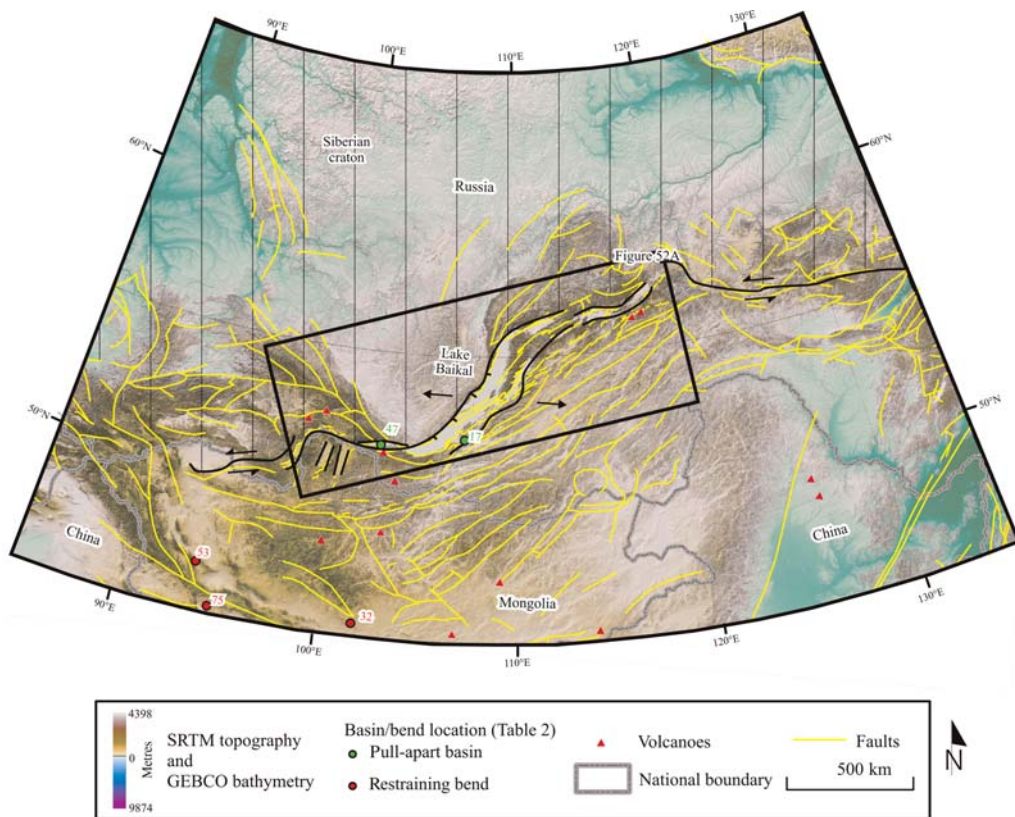


Fig. 51. Tectonic setting of the Lake Baikal pull-apart basin in eastern Russia. Faults are from a compilation by Howard *et al.* (2003), and Levine (1995).

fault zone), to explain their observations. Another active trench-linked fault in southern Chile, the right-lateral Liquine–Ofqui strike-slip fault zone, has the correct shear sense for the direction of oblique subduction (Thomson 2002).

Bends. As in the case of Sumatra, the strike-slip and normal faults of the Chilean margin closely follow the changing trend of the Chile Trench, although the trace of the Atacama fault zone has a more sinusoidal trace than the trench itself (Fig. 58). The Cerro de Mica lazy-Z restraining bend is formed at an 8-km-wide fault separation on two faults adjacent to the present-day trace of the Atacama fault zone (Fig. 58). The internal structure of the Cerro de Mica is comparable with laboratory bends of the same type created by McClay & Bonora (2001). No pull-aparts have been reported from the Atacama fault zone.

The Salina de Fraile lazy-S pull-apart (Reijs & McClay 2003) formed on a left-lateral fault in the Argentinian Andes, presumably as a response to the

same tectonic mechanism for left-lateral faulting proposed by Armijo & Thiele (1990; Fig. 59a). The basin exhibits oblique normal faults at its edge, and a cross-basin strike-slip fault zone similar to those described in young pull-aparts by Mann *et al.* (1983) and Zhang *et al.* (1989), and to those produced by Dooley & McClay (1997) in laboratory experiments.

Tectonic setting and geological development of active bends of the Kurile Islands strike-slip fault system, Kurile Islands and northern Japan

Kimura (1986) first proposed the existence of a fore-arc sliver adjacent to the southern Kurile Arc that is driven by the oblique subduction of Pacific crust (Fig. 60). Using data from both onland and offshore, he described a submarine rift at the NE end of the sliver, and a small fold–thrust belt that extended on land into northern Japan (Hokkaido Island; Fig. 61). DeMets (1992) used earthquake focal

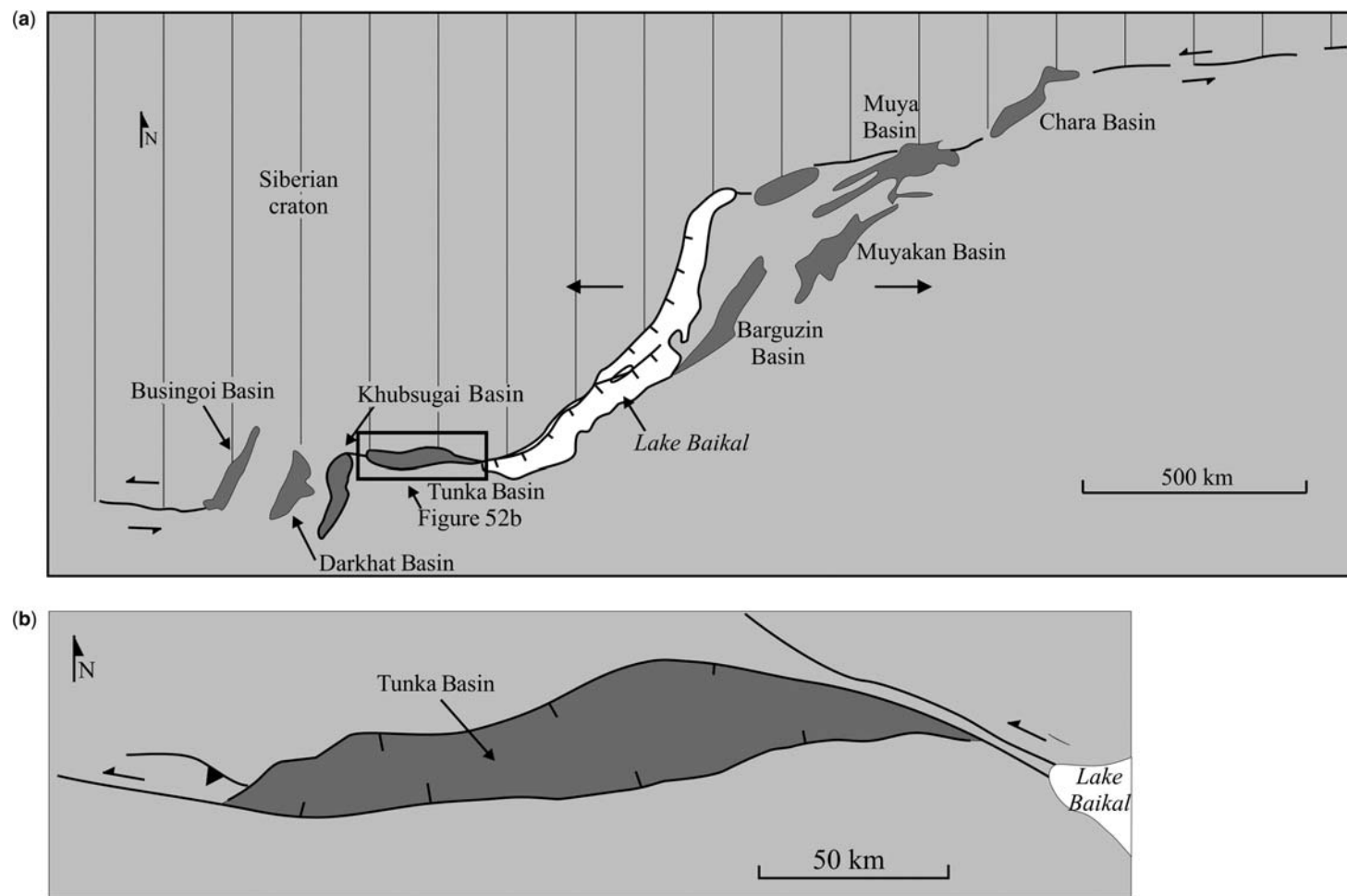


Fig. 52. (a) Schematic map of the Lake Baikal pull-apart basin, modified from Larroque *et al.* (1999). (b) Schematic map of the Tunka pull-apart basin, modified from Larroque *et al.* (1999).

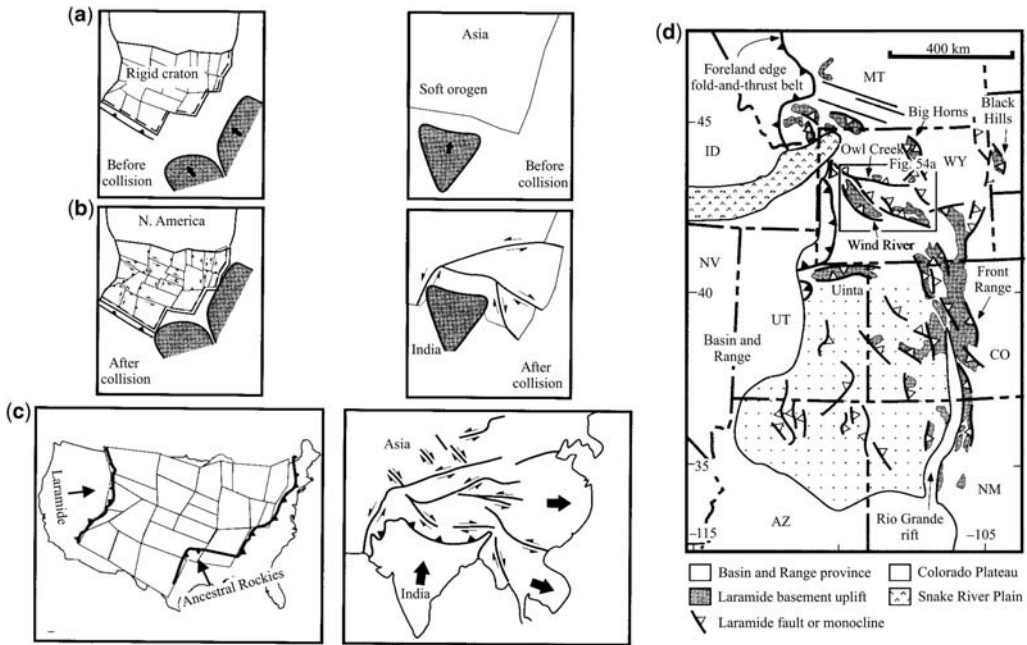


Fig. 53. (a) Comparison maps from Marshak *et al.* (2003) of the interior platform of the United States at the time of the Alleghenian–Ouachita Orogeny (left), and the eastern Eurasia Orogeny produced by indentation by the Indian continent (right). The interior platform of the US is a rigid craton whose upper crust has been broken into a mosaic of blocks by faults. In eastern Eurasia, the southern margin of the continent is a soft Phanerozoic orogen. (b) During the Alleghenian–Ouachita collision, the craton was strained only slightly, so that the internal crustal blocks moved only slightly. In eastern Eurasia, crustal blocks underwent eastward lateral escape as the lithosphere was strained significantly. (c) The mosaic of crustal blocks in the interior platform of the US contrasts with the major regional faults in eastern Eurasia. (d) Map of the US, showing the locations of the Alleghenian–Ouachita Orogeny responsible for the Palaeozoic Ancestral Rockies, and the Laramide Orogeny responsible for reactivation of the previously formed Ancestral Rockies along with formation of new structures in the Late Cretaceous and Early Cenozoic. (d) Regional map from Marshak *et al.* (2003), showing the distribution of Laramide basement uplifts and structures of the Colorado Plateau and the Rocky Mountains.

mechanisms to support the existence of the Kurile fore-arc sliver and a 6–11 mm/a rate for its SW translation relative to the North American plate.

Bends. As in the case of the Sumatran Fault, the unnamed left-lateral strike-slip fault bounding the sliver curves strongly into the rifted area and may indicate enhanced crustal thinning within the accretionary wedge (Fig. 61). One rhomboidal restraining bend is shown by Kimura (1986) along this fault adjacent to the Kurile Islands of Etoroph and Urup.

Tectonic setting and geological development of active bends on strike-slip faults in Japan

Oblique subduction beneath the central part of Japan induces left-lateral faulting along the

Median Tectonic Line, the longest strike-slip fault in Japan (Fig. 62). The Median Tectonic Line formed along a terrane boundary within the crust, and gentle restraining bends in the fault trace are marked by elongate outcrops of blueschist rocks (Mann & Gordon 1996). Part of this shortening is accommodated by an extensive zone of right-lateral faults, which are much shorter in length but distributed over a larger area (Fig. 62).

Bends. The 1995 Kobe earthquake – which, at a cost in excess of 200 billion US dollars, was one of the most expensive natural disasters in history – ruptured a zone that is subparallel to the Median Tectonic Line (Fig. 62). Aftershocks outline a rhomboidal pull-apart along this fault near the epicentral area (Sato *et al.* 1998). The adjacent Rokko Range uplift may be a restraining bend paired to this releasing bend (Sangawa 1986).

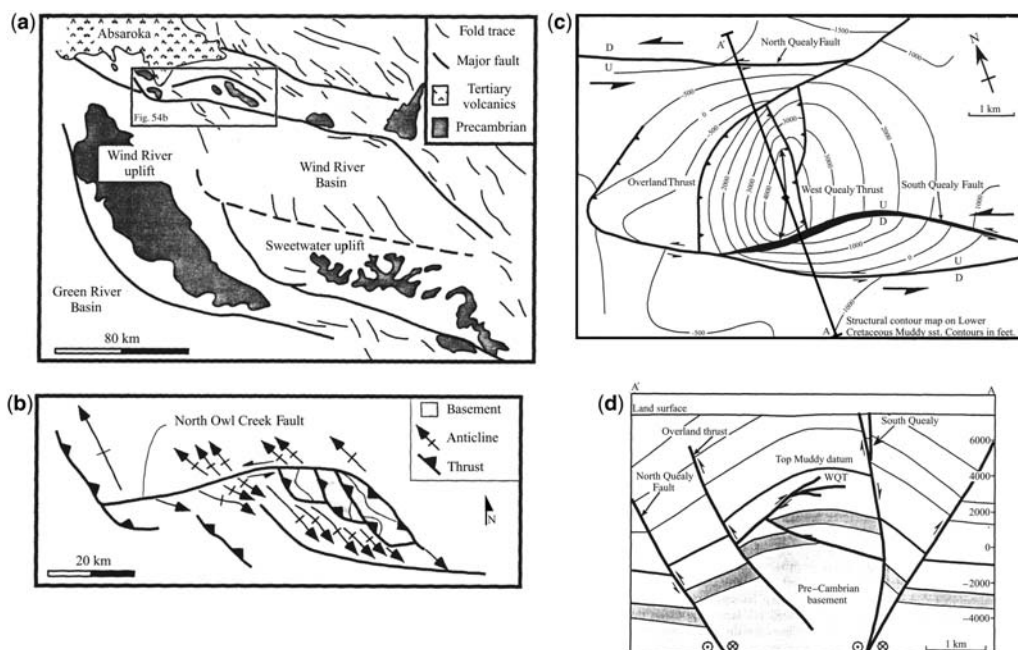


Fig. 54. (a) Map from Marshak *et al.* (2003), redrawn from Stone (1969), showing en échelon folds suggestive of strike-slip faulting through Wyoming formed during the Laramide Orogeny. (b) Map from Marshak *et al.* (2003), redrawn from Paylor & Yin (1993), showing en échelon folds in a restraining bend along the North Owl Creek Fault in Wyoming. (c) Map from Stone (1995) in McClay & Bonora (2001), showing the Quealy restraining bend in the Laramie Basin, Wyoming. (d) Cross-section through the Quealy restraining bend, from Stone (1995) in McClay & Bonora (2001).

Tectonic setting and geological development of active bends along the Central American fore-arc sliver

The presence of strike-slip faults along the volcanic line of the Central America Arc, as a possible indicator of the edge of a fore-arc sliver, was first proposed on the basis of earthquakes, by Harlow & White (1985; Fig. 63a) and White & Harlow (1993). DeMets (2001) placed kinematic constraints on the fore-arc sliver using earthquake focal mechanisms, and estimated a rate of 9 mm/a for the northward displacement of the sliver. Unlike Sumatra, the Kuriles and Japan, the fore-arc sliver is driven by less than 10° of oblique plate convergence.

Bends. For many years, there was no geological evidence for active strike-slip faults parallel the volcanic arc. Carr (1976) and Corti *et al.* (2005) identified active right-lateral faults in the El Salvador strike-slip fault system that roughly parallels the volcanic arc. The fault zone contains one prominent

rhomboidal pull-apart Basin, the Rio Lempa Basin (Fig. 63b). To the south, in Nicaragua, Girard & van Wyk de Vries (2005) have proposed similar pull-apart basins, although these features are largely covered and have not been mapped in the subsurface.

Tectonic setting and geological development of active bends of the Philippine fault system

Introduction. The 1200-km-long, left-lateral Philippine strike-slip fault system is a trench-linked strike-slip fault related to oblique subduction of the Philippine Sea plate beneath Eurasia (Fitch 1972; Barrier *et al.* 1991; Fig. 64). Motion at a rate of 1–5 mm/a on the Philippine fault zone and Philippine Trench occurred in the Late Neogene as a back-thrusting response to underthrusting of the Palawan Block along the western edge of the block.

Bends. The fault has a pronounced arcuate shape that is convex to the NE and may reflect the

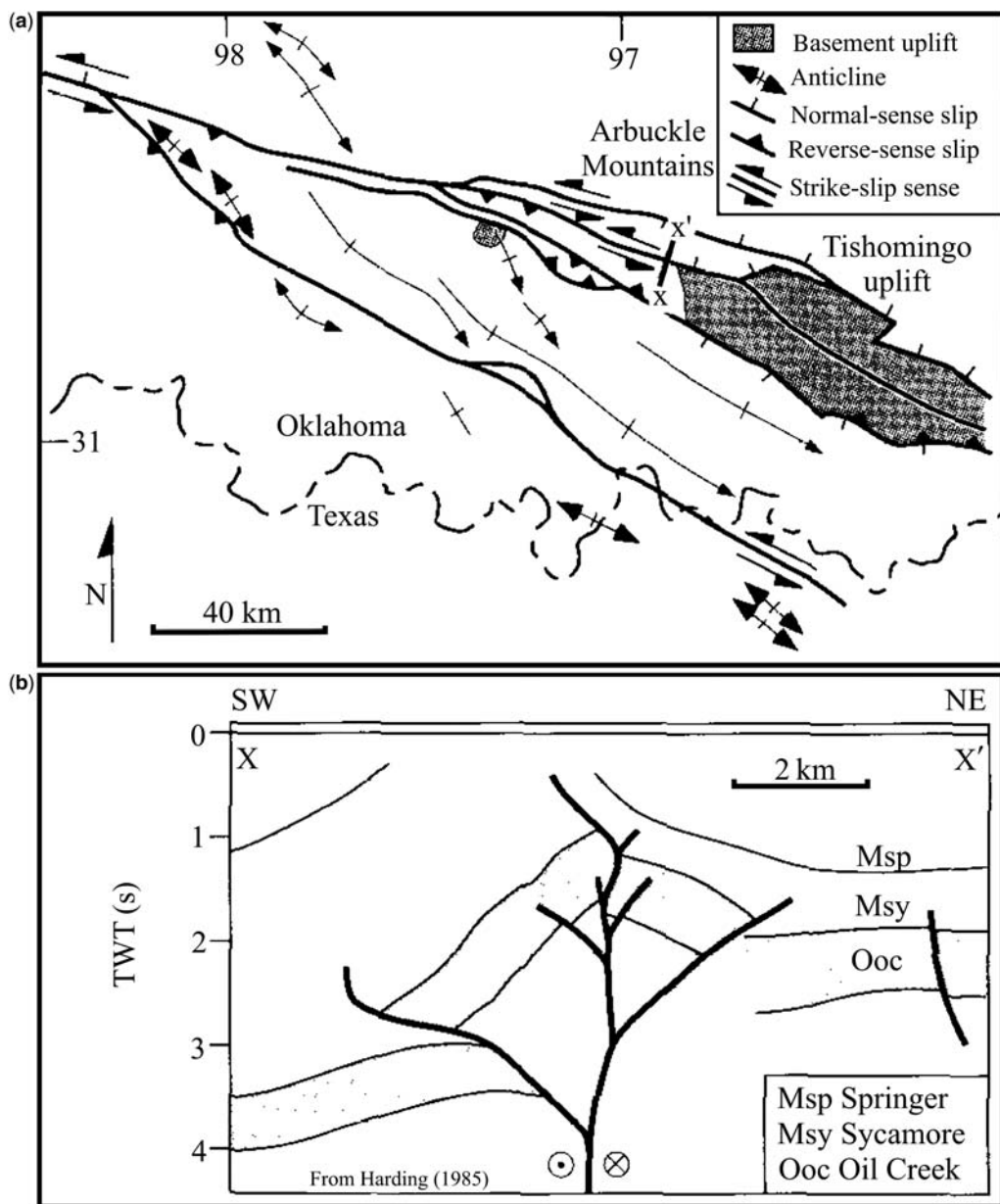


Fig. 55. (a) Map of part of the southern Oklahoma aulacogen, including the Arbuckle Mountains and the Ardmore Basin. (b) Flower-structure interpretation of a seismic line across the Ardmore Basin (at X–X'). Modified from Harding (1985).

collision of the Palawan Block (Fig. 64). Geological studies (Barrier *et al.* 1991) and GPS studies (Aurelio 2000) show that motion along this curved part of the fault is almost pure strike-slip. The northern terminus of the fault is characterized

by an apparent large restraining-bend structure (Fig. 64). An inverted releasing bend containing Late Miocene–Early Pliocene sedimentary rocks has been described from the northern end of the fault system (Ringebach *et al.* 1990).

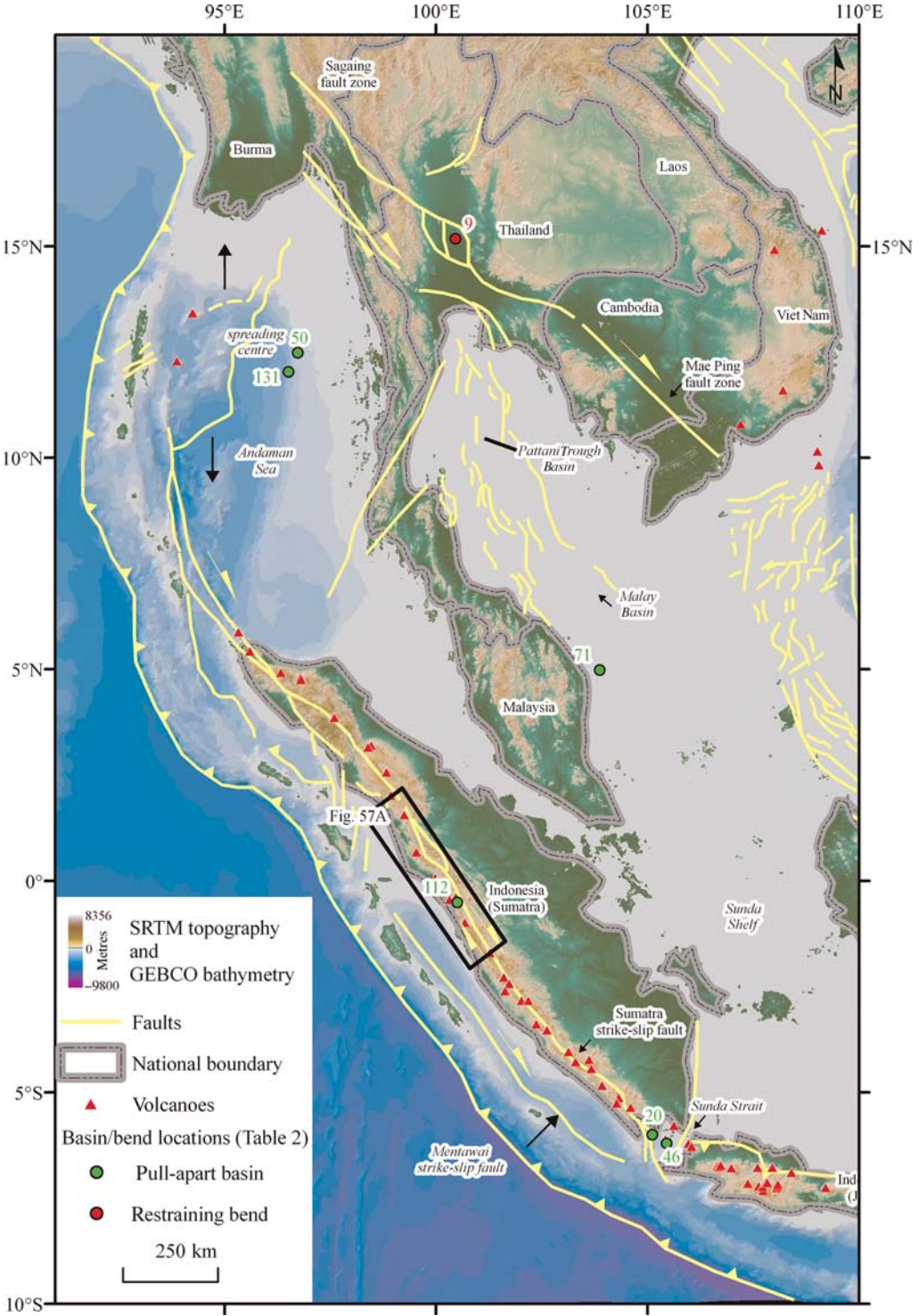


Fig. 56. Regional tectonic map of active to Cenozoic strike-slip fault systems of Sumatra and SE Asia, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilation by Levine (1995), and Coffin *et al.* (1997). Box shows the location of the detailed map in Figure 57.

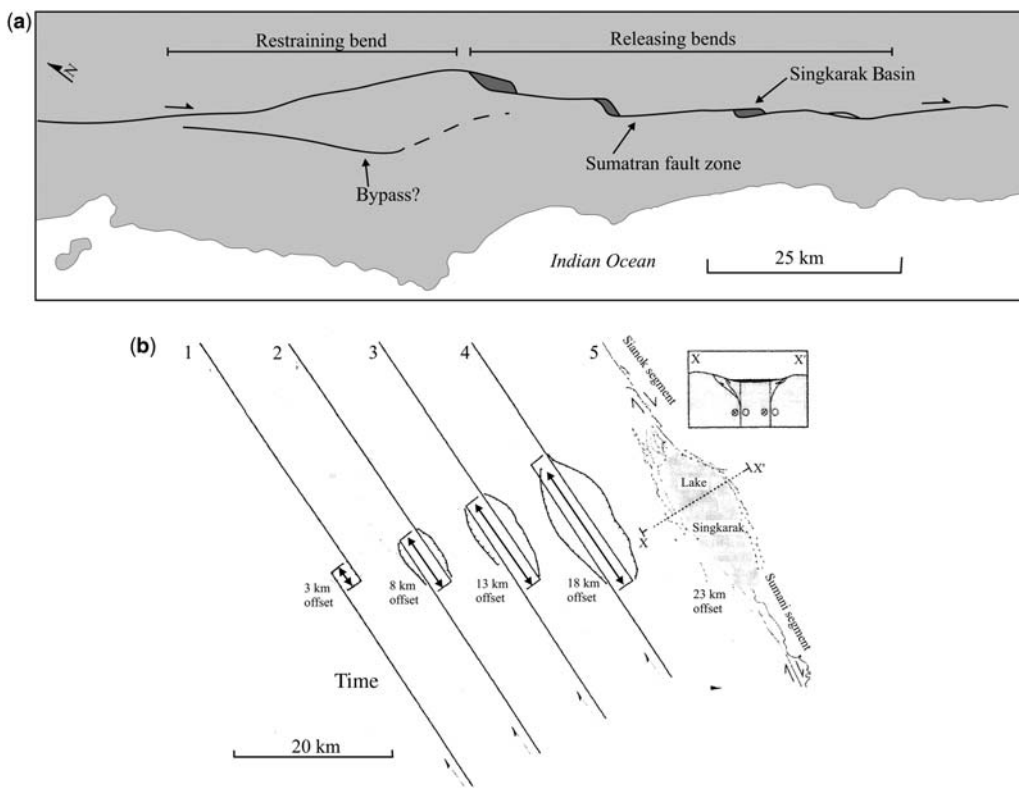


Fig. 57. (a) Sketch map of a large paired-bend structure along the Sumatran fault zone, from mapping by Sieh & Natawidjaja (2000). (b) Evolutionary model by Sieh & Natawidjaja (2000), showing the progressive opening of the Lake Singkarak pull-apart basin along the Sumatran fault zone.

Tectonic setting and geological development of ancient bends of the Berengian margin

The Berengian margin, or the eastern margin of the Bering Sea adjacent to Alaska, was mapped by the oil industry during routine hydrocarbon exploration in the 1970s and 1980s (Worrall 1991). This wealth of subsurface seismic-reflection and drill data revealed a 1000-km-long, right-lateral strike-slip system active from during the Palaeocene and Eocene (Fig. 65). As in Sumatra and other modern arcs, this margin was characterized by a strike-slip fault parallel to a subduction zone and defined a fore-arc sliver.

Bends. Subsurface mapping has shown several excellent examples of paired-bend geometries (St George pull-apart basin and Black Hills restraining bend, Fig. 66a) and isolated rhomboidal pull-aparts (Fig. 66b) and sharp restraining bends (Fig. 66c).

Tectonic setting and description of bends on cratonic strike-slip faults

I am adding a fifth category of bends to the other classes proposed by Woodcock (1986; Table 1). I call this fifth class the 'bends on cratonic strike-slip faults'. These are defined as bends in cratonic areas that are far removed from active plate boundaries, yet are still subject to intraplate stresses. These bends, which are remarkably similar to sharp restraining bends, are defined mainly on the basis of their intraplate earthquake activity, and in many cases seismically reactivate known basement structures that generally correspond with failed rifts (Gangopadhyay & Talwani 2003). Cratonic settings of strike-slip faults differ from the areas of Central Asia discussed by Cunningham (2007) because the Central Asian examples that are all found in mechanically weak Palaeozoic accretionary belts that separate areas of old, cold cratonic areas and are subject to distant stresses induced by continental collisions.

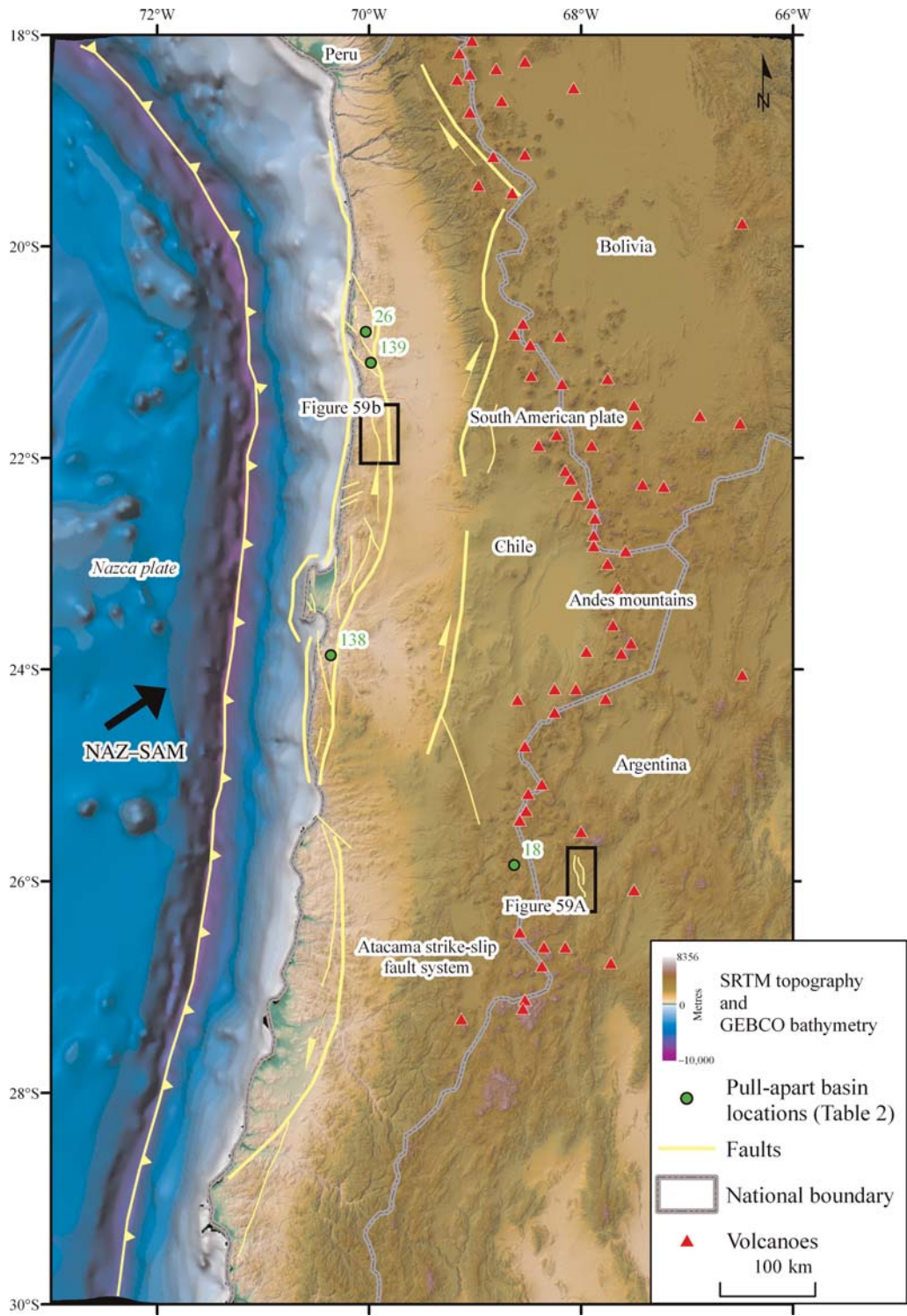


Fig. 58. Regional tectonic map of active to Cenozoic strike-slip fault systems of Chile, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults taken from compilations by Levine (1995) and Armijo & Thiele (1990). Boxes show locations of detailed maps in Figures 59A & 58B.

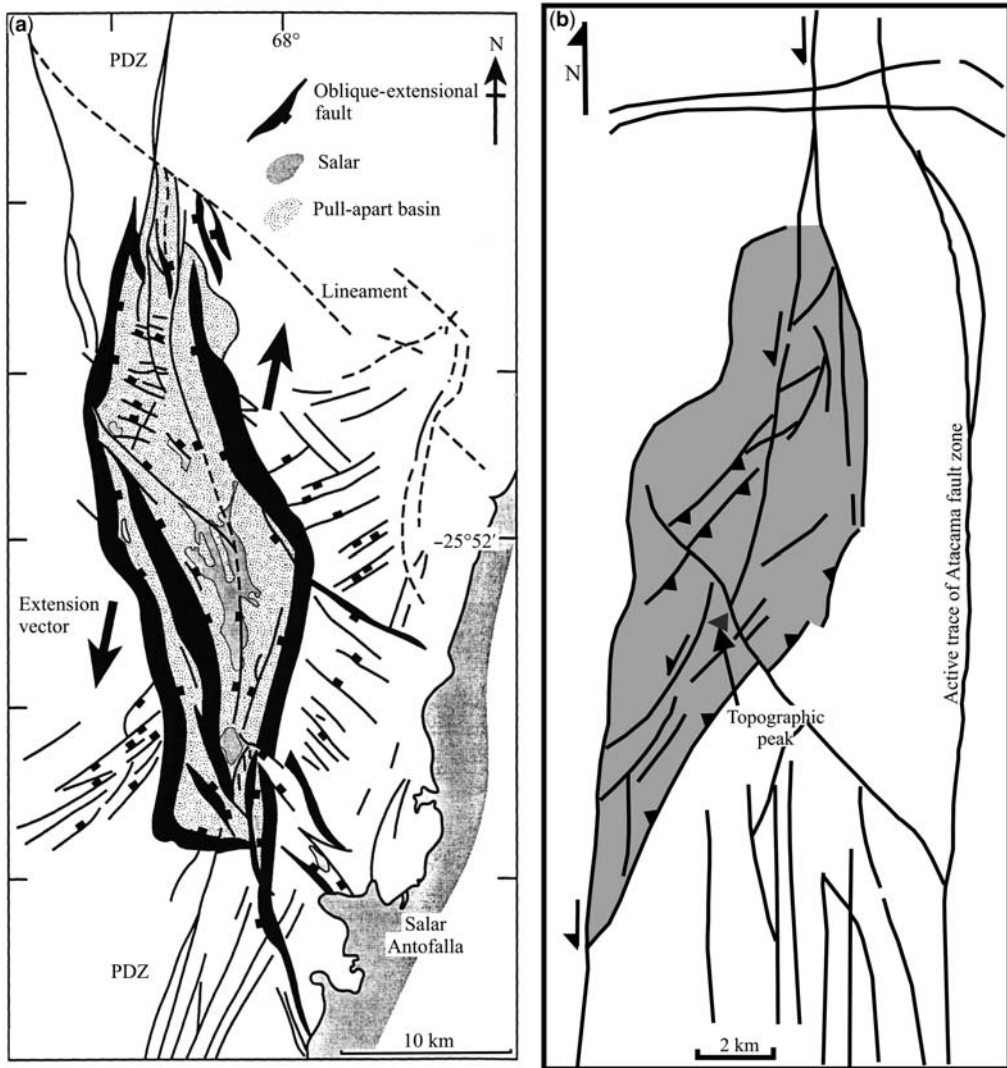


Fig. 59. (a) Map of the Salina del Fraile releasing bend in the northern Argentinian Andes, from Dooley & McClay (1997). (b) Map of the Cerro de la Mica restraining bend along the Atacama fault system in northern Chile (from McClay & Bonora 2001).

New Madrid area, USA. This is the most seismically active part of the mid-continent of North America and was the site of three or four large earthquakes in the period of 1811–1812, all with a body-magnitude of 7.0 or greater (Marshak *et al.* 2003). Earthquake epicentres define a bend in a fault that has reactivated older faults in the underlying Reelfoot Rift, a NNE-trending failed rift into continental crust of North America. The rift formed in Late Proterozoic to Early Cambrian times (Fig. 67a). Focal mechanisms indicate that earthquakes on the NE and SW fault segments are

right-lateral strike-slip, and that earthquakes on the central segment are thrusts with the NE side up (Gangopadhyay & Talwani 2003). The deformation is consistent with reactivation of the ancient faults by the present-day NE–SW-trending maximum compressive stress trajectory of the central part of the North American continent (Marshak *et al.* 2003; Fig. 67a).

Middleton Place–Summerville seismic zone, South Carolina. This cluster of seismicity was the

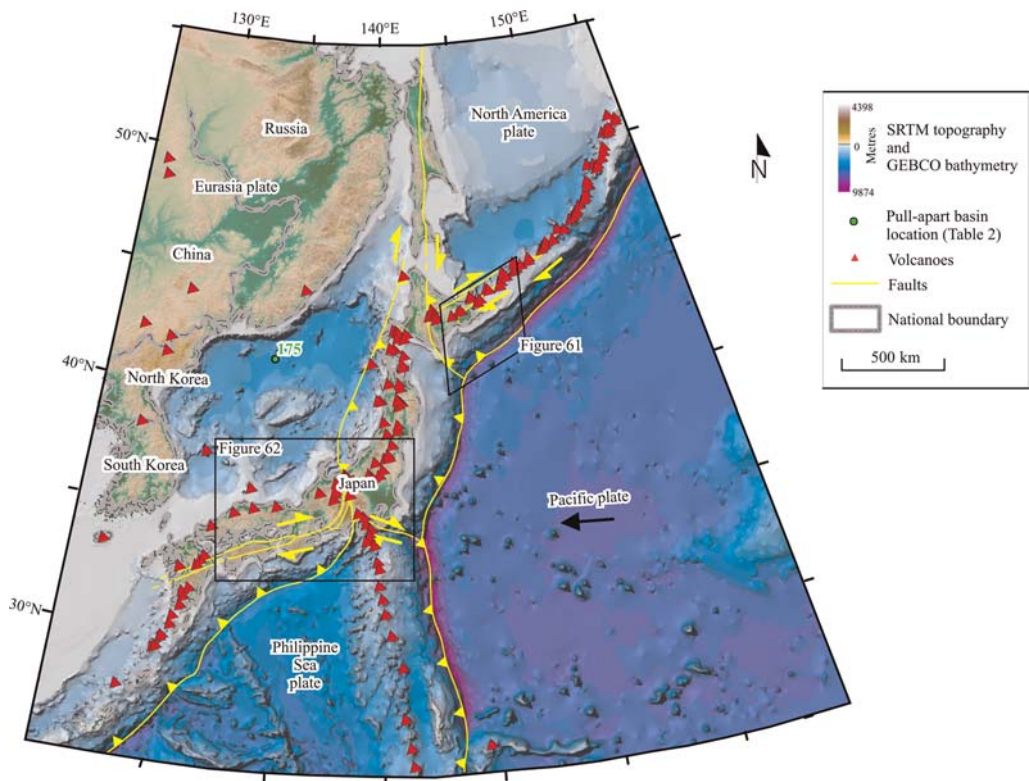


Fig. 60. Regional tectonic map of active to Cenozoic strike-slip fault systems of Japan, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults, taken from compilations by Mann & Gordon (1996) and Coffin *et al.* (1997). Boxes show the locations of detailed maps in Figures 61 & 62.

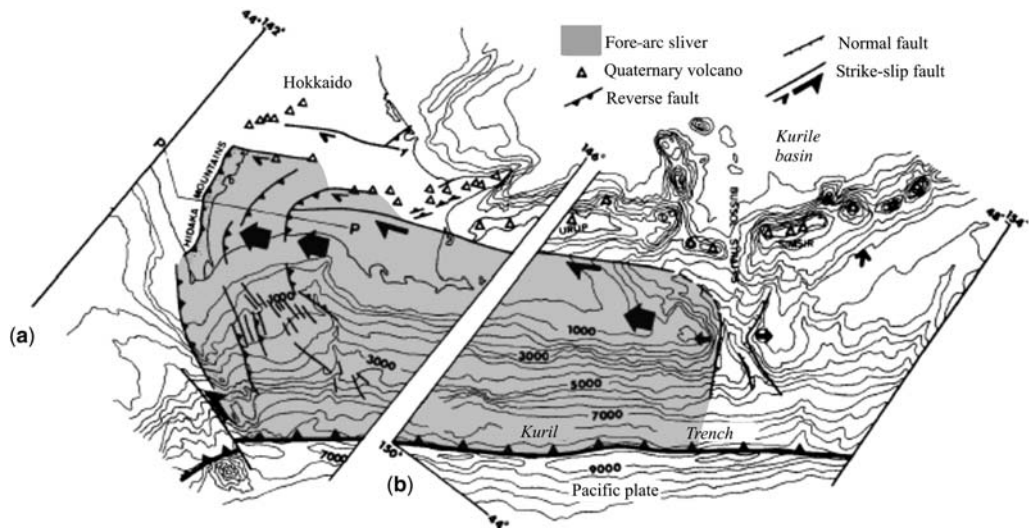


Fig. 61. Active fore-arc sliver (shaded in grey) along the SW Kurile Arc north of Japan, from Kimura (1986). Oblique subduction of the Pacific plate beneath the Kurile arc drives the sliver to the SW into a mini-collision zone with the northern island of Japan (Hokkaido). An unnamed active strike-slip fault connects the extensional zone in the Bussol Strait with a collisional zone on and near the island of Hokkaido in northern Japan.

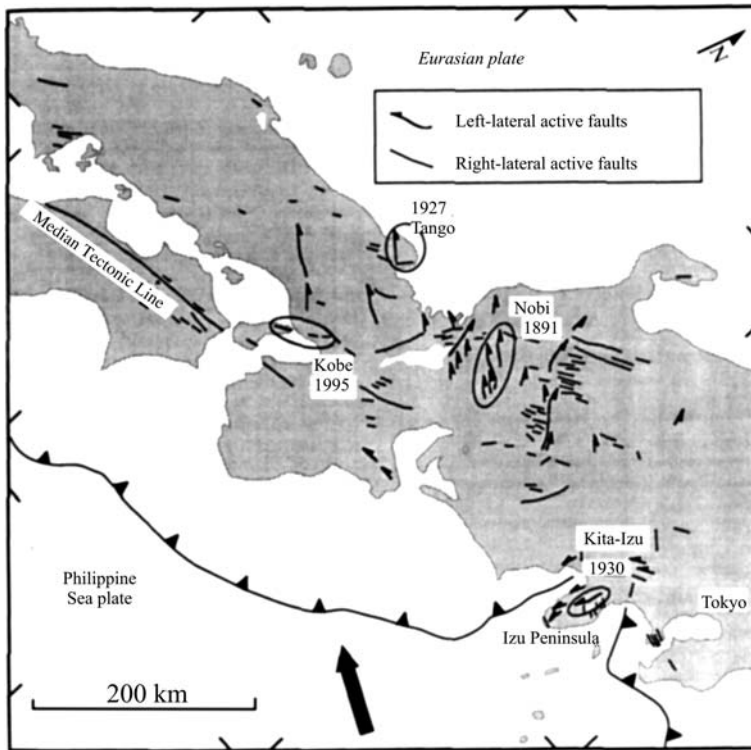


Fig. 62. Many active faults in Japan are driven by oblique subduction right-lateral displacement along the Median Tectonic Line (modified from Yeats *et al.* (1997). Ellipses show the locations of major damaging shallow strike-slip earthquakes related to deformation of the overriding plate in Japan, including the fault responsible for the 1995 Kobe earthquake, one of the most economically devastating earthquakes in history.

source of the 1886 Charleston (South Carolina) earthquake, and is currently the most active zone of seismicity in the South Carolina coastal plain (Gangopadhyay & Talwani 2005; Fig. 67B). As in the case of the New Madrid seismic zone, this cluster overlies a junction of faults related to the Mesozoic South Georgia rift basin. Earthquake focal mechanisms and modelling show that the NE and SW fault segments are right-lateral strike-slip, and that earthquakes on the central segment are thrusts with the NE side up (Gangopadhyay & Talwani 2005). The present-day maximum compressive stress trajectory of the eastern part of the North American continent is NE–SW and consistent with this kinematic interpretation.

Significance and relevance to strike-slip bends.

Both bends have been explained in terms of the pre-existing faults acting as ‘stress concentrators,’ which cause anomalous buildups in their vicinity within a localized volume of weak crust (Gangopadhyay & Talwani 2005). Gangopadhyay & Talwani (2003) have compiled 39 examples of

similar tectonic settings of seismic areas at fault intersections mostly within failed rift structures on continental margins far from the effects of active plate boundaries. One fault is typically longer than the other, and is favourably oriented for failure—given the direction of the maximum compressive stress. The recurrence interval of major earthquakes on these zones is much longer than similar faults in active plate boundaries.

Discussion

A major point of this lengthy introduction was to show that the structural styles of about 200 well-documented releasing and restraining bends are remarkably similar, despite the diverse nomenclature that has arisen to describe them and their different strike-slip settings (Fig. 2, Table 1). Releasing bends in all settings far outnumber restraining bends. In this discussion, I will point out a few of the major differences seen in bends formed in the four tectonic settings; speculate on

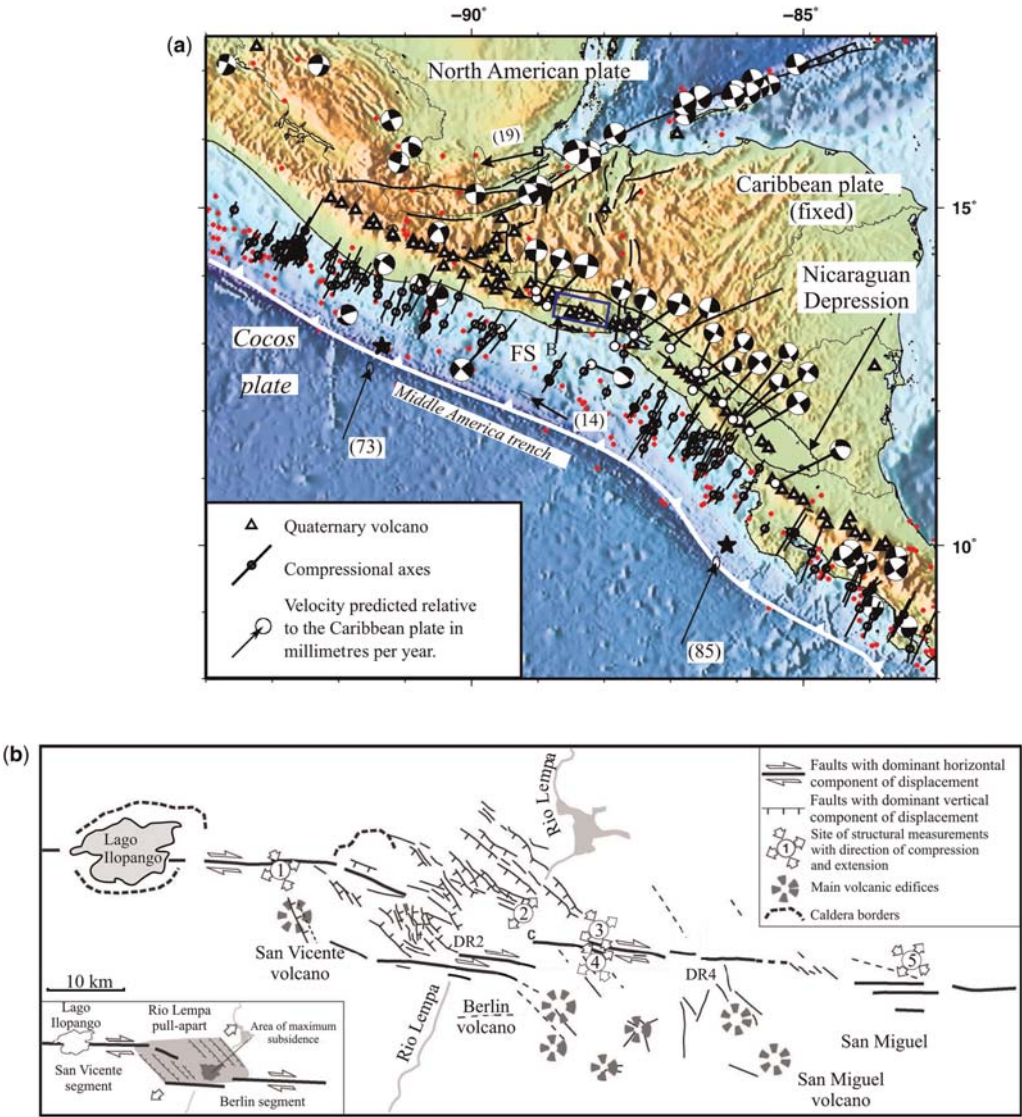


Fig. 63. (a) Regional tectonic map of active to Cenozoic strike-slip fault systems of Central America, using SRTM topography and ETOPO bathymetry as a base map. Black lines are active faults, taken from a compilation by DeMets (2001). (b) Schematic structural map of interpreted pull-apart by Corti *et al.* (2005).

their tectonic origins; and suggest some possible areas for future work.

Pull-apart development

The geological model proposed by Crowell (1974a, b) and modified by Mann *et al.* (1983) has been applied with success to many releasing bends worldwide (Fig. 7). The largest problem area is why do the strike-slip faults become curved in the

first place for the releasing bend to form? In continental-boundary settings, many pull-aparts occur as part of a paired bend. Paired bends or fault zone ripout structures may form by a process of fault abandonment and reoccupation, as outlined by Swanson (2005; Fig. 12b). One open question is whether the restraining bend or the releasing bend appears first to form a paired bend. My proposal in this paper and in Mann *et al.* (2007a) is that the restraining bend forms first as a result of the

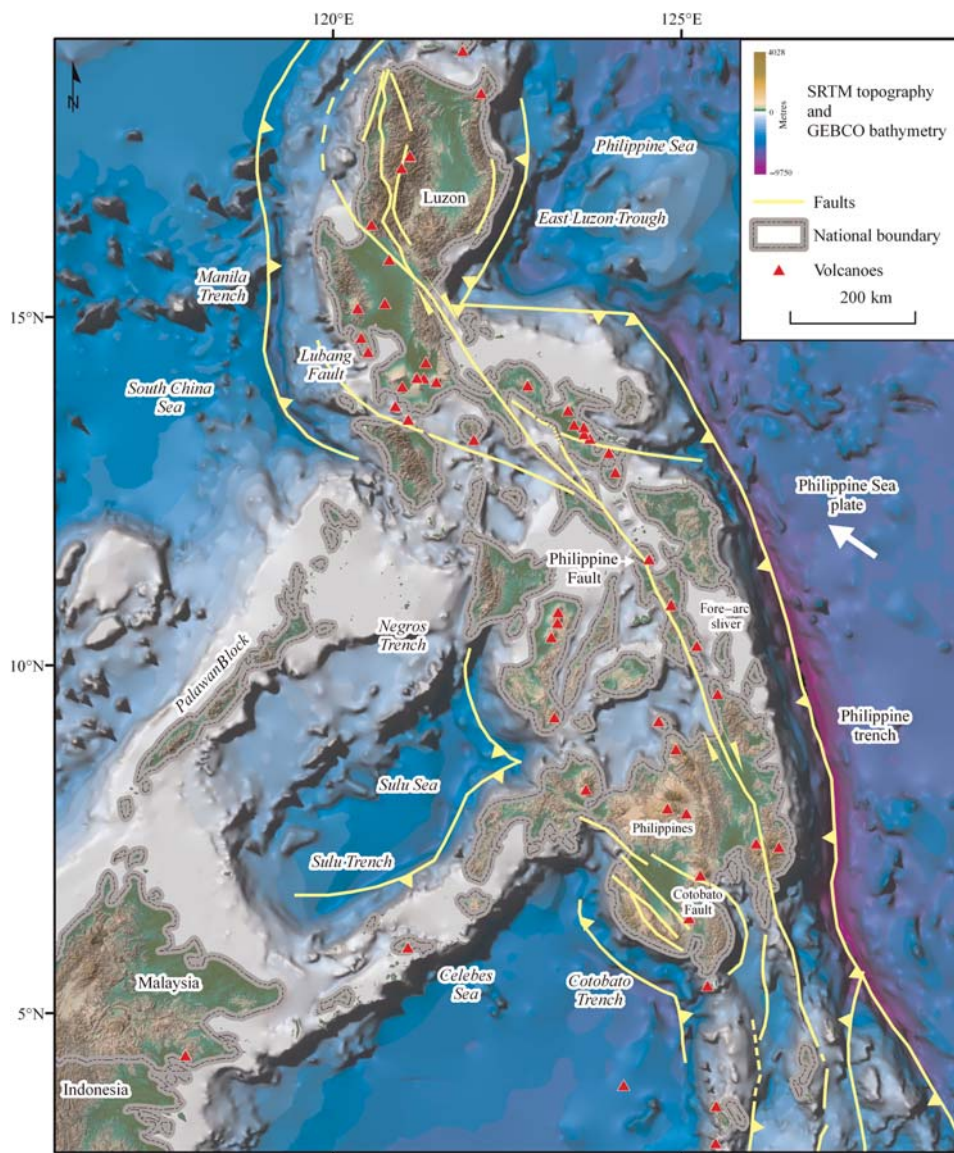


Fig. 64. Regional tectonic map of active to Cenozoic strike-slip fault systems of the Philippines archipelago, using SRTM topography and GEBCO bathymetry as a base map. Yellow lines are active faults, taken from a compilation by Coffin *et al.* (1997). GPS vector taken from Barrier *et al.* (1991).

interaction of strike-slip faults with a pre-existing basement feature. This proposal could be tested by careful dating of the initiation of adjacent pull-aparts and restraining bends that make up a paired bend.

Another pull-apart problem is why do basins become extinct by developing a cross-basin fault (Zhang *et al.* 1989)? One idea is that as faults

adjust to changing plate kinematics, new faults may form to allow the fault to remain in a slip-parallel orientation (Fig. 8a). Another idea is that bends on faults can form as a natural consequence of shear-zone evolution, but this model would only apply to new faults with little offset (Fig. 12a).

Tectonic settings appear to account for few differences in the style of opening of pull-apart

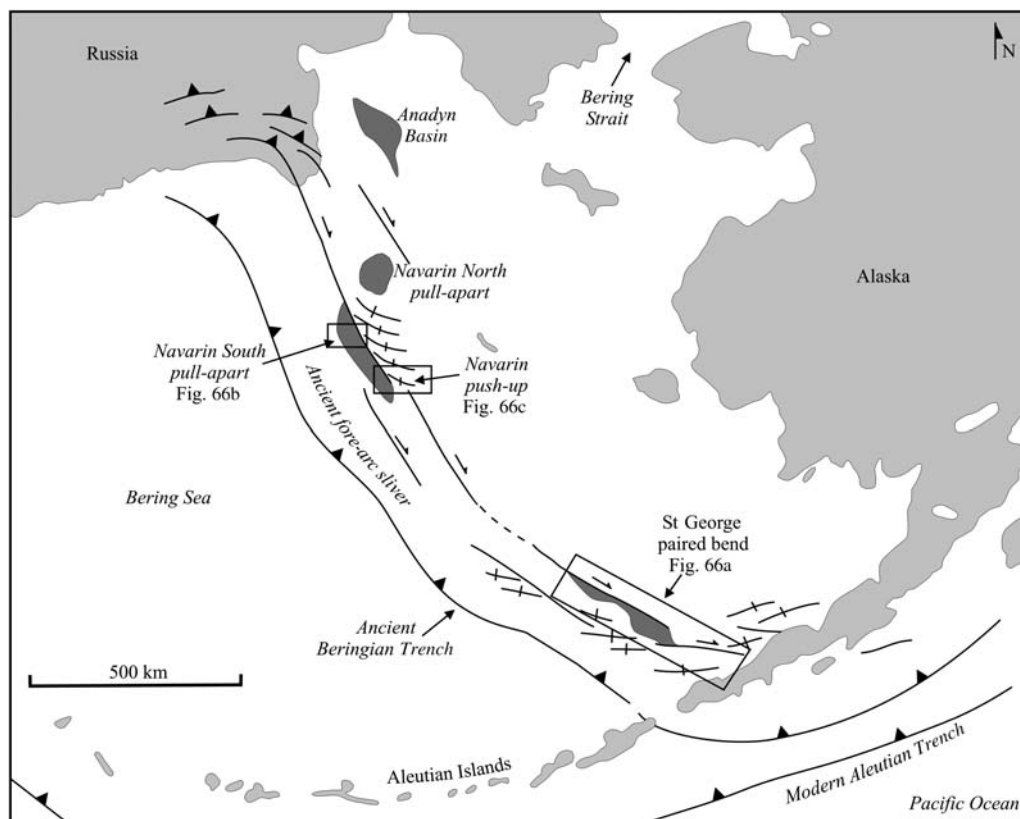


Fig. 65. Sketch map from Worrall (1991), showing Palaeogene strike-slip faults and bends along the Bering Sea margin.

basins. Larger basins appear to reflect larger angular differences between the original curved trend of the fault and the plate vector. Pull-aparts appear much less common in indent-linked systems like the Himalayas and the Tibetan Plateau (Fig. 47a), and are more common on long continental boundaries like the San Andreas (Fig. 15) and Caribbean faults (Fig. 23).

Restraining-bend development

Restraining-bend models are far less evolved and tested than those for releasing bends. The choices include the shear-zone model that would be relevant only for a newly formed fault; the sidewall ripout model which Swanson (2005) has shown to be relevant to many natural fault settings independent of their scale; and the paired-bend model proposed here.

In the paired-bend model, the formation of a restraining bend is followed by the formation of an adjacent releasing bend; the releasing bend

forms as the strike-slip fault readjusts to its bent configuration (Fig. 8c). I speculate that the final stage in this process is the formation of a 'shortcut' or 'bypass' fault, which effectively bypasses the area of active bend deformation and returns the fault to an original, or at least straighter orientation. The reason why the strike-slip fault returns to its original, straight trace after being diverted at the bend is not clear. One explanation is that it is easier for the fault to reoccupy a zone of weakness parallel to the interplate slip vector (that is, its former fault trace) than breaking a new fault. A second explanation is that increased convergence at the restraining bend causes the paired-bend area to lock and be 'lopped off' by the more efficient bypass fault. A third explanation is based on the conservation of crustal volume in the bends area: crustal thickening in the restraining-bend area is compensated by thinning in the adjacent pull-apart area in order to maintain a crustal balance.

The paired-bend model is consistent with one of the fundamental differences between continental

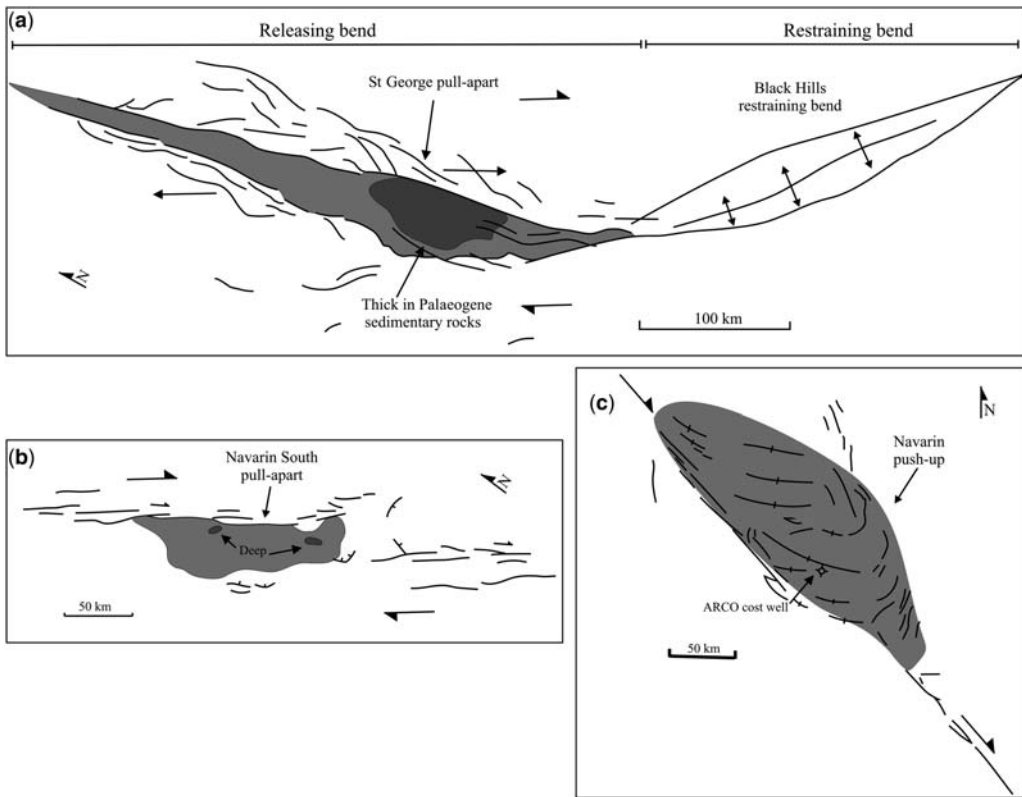


Fig. 66. Sketch maps modified from maps by Worrall (1991), showing details of the Bering Sea strike-slip zone. Location shown on Figure 65. (a) Paired-bend geometry of the St George pull-apart basin and the Black Hills restraining bend. (b) Navarin South pull-apart basin. (c) Navarin restraining bend.

strike-slip faults and oceanic transform faults: the rarity of restraining bends along the main fault trace, or 'principal displacement zone' of oceanic transform faults (Garfunkel 1986). I attribute the rarity of restraining bends along oceanic transform faults to the lack of pre-existing crustal heterogeneities in the oceanic crust on which restraining fault bends nucleate. Oceanic transform faults are instead characterized by divergent fault structures (for example, pull-apart basins, wide fault valleys) which probably form as a direct consequence of changes in relative plate motion (Garfunkel 1986; Fig. 8a).

This process of restraining-bend formation, uplift, rapid erosion, fault readjustment, and releasing-bend formation appears to be the dominant tectonic mechanism for active deformation and fault-controlled sedimentation along many of the fault systems reviewed here and may, therefore, provide a predictive model for interpreting other active and ancient strike-slip plate boundaries. The key to proving this hypothesis lies in precise age

determinations of the initial age of uplift of the restraining bend and the age of initial subsidence of the releasing bend. Because active restraining bends generally form land areas, detailed geological mapping and stratigraphic studies are needed to determine their uplift history. Marine seismic surveys and deep-sea drilling are needed to determine the history of adjacent releasing bends, which commonly occur in deep-water areas.

Conclusions

The main conclusions and questions raised by this compilation include the following:

1. There are systematic structural relationships between releasing and restraining-bend segments along active strike-slip faults. Previous workers have not emphasized synchronous tectonic activity at paired restraining and releasing bends, because paired bends commonly

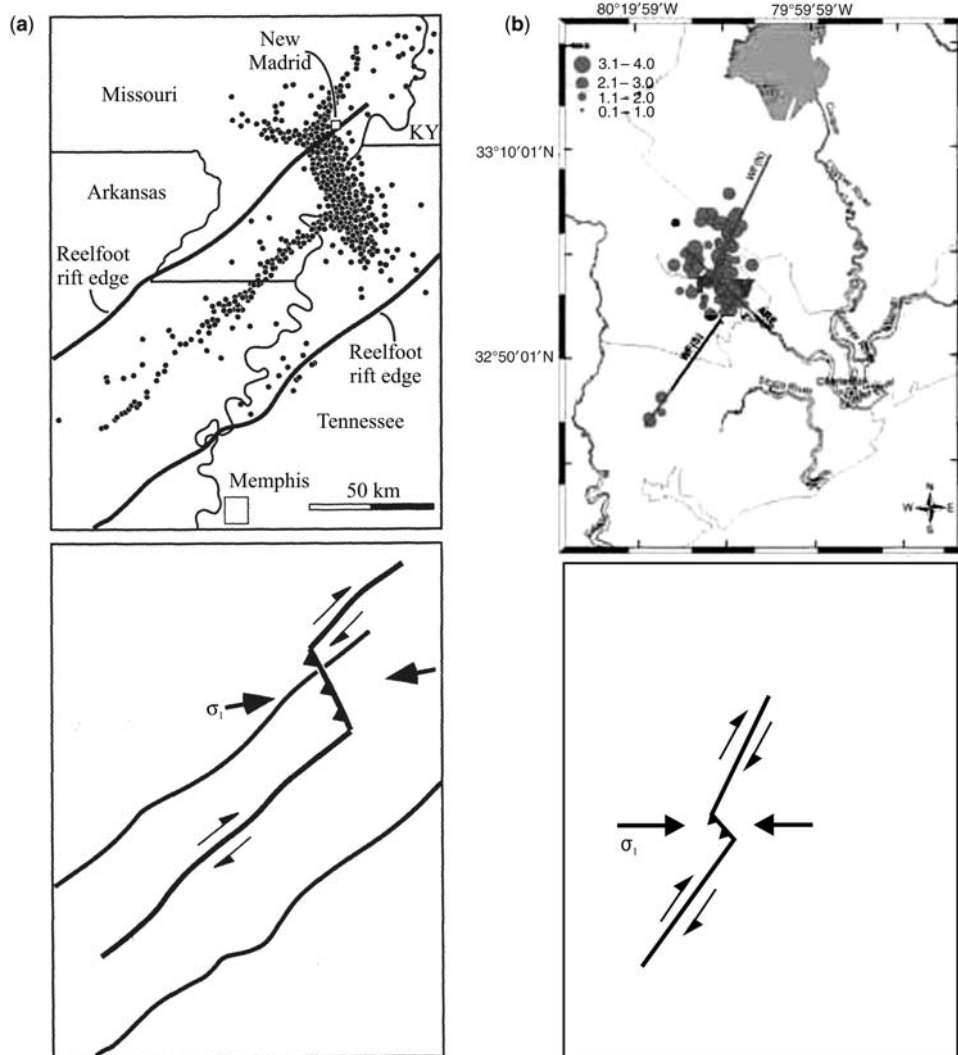


Fig. 67. (a) Map showing the location of earthquake epicentres in the New Madrid seismic zone in southern Missouri. Sketch showing restraining-bend structure based on fault-plane solutions (from Marshak *et al.* 2003). (b) Map showing the location of earthquake epicentres in the Middleton Place Summerville seismic zone (from Gangopadhyay & Talwani 2005). Sketch showing restraining-bend structure based on fault-plane solutions (modified from Gangopadhyay & Talwani 2005).

- extend over land and sea areas with along-strike lengths of tens to hundreds of kilometres.
- In paired-bend systems, some restraining bends appear to originate first as a result of interference between the strike-slip faults and pre-existing structures in the basement. In the case of Jamaica and Lebanon, pre-existing structures correspond with extinct rifts that strike at an angle of about 45° to the strike-slip fault.
- The rarity of restraining bends along oceanic transform faults probably reflects the lack of pre-existing crustal heterogeneities in oceanic crust.
- Which comes first: the restraining or releasing bend? The formation of two major restraining bends (Jamaica, Lebanon) at the sites of extinct rifts suggests that bends may originate first, as a result of strike-slip reactivation.

5. Why does the strike-slip fault crossing a paired bend return to its original trace after being diverted at the bend? One explanation is that it is easier for the fault to reoccupy a zone of weakness parallel to the interplate slip vector (that is, the old fault zone) than breaking a new fault through an unfaulted area. A second explanation is that increased convergence at the restraining bend causes the paired-bend area to lock and be 'lopped off' by the more efficient bypass fault.
6. Is the uplift history of the restraining-bend area consistent with the subsidence history of the releasing bend? In many areas, there are not enough data from either the releasing- or restraining-fault area to show clearly that the two areas deformed simultaneously. The paired-bend model is testable in active strike-slip zones by combined marine and land studies capable of precise age determinations of the time of uplift of the restraining bend and the time of subsidence of its paired releasing bend.

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