CHAPTER 5.
SYNTHESIS: STRUCTURAL-TECTONIC MODELS AND TIMING RECONSTRUCTION

In the last chapter of this first part of the study, we combine the theoretical considerations made in the previous chapters (about the applicability and limitations of the applied methods to the study of active tectonics in mountainous intracontinental settings and related basin formation) with the state-of-the-art of the conception of tectonism related to continental deformation. The idea's explained and elaborated in this chapter will be of use for the perception and development of the reasoning for interpreting the studied areas, discussed in the next part of this study.

5.1 Introduction

In the first part of the study we discussed the tools used for studying brittle tectonics in active regions, relative fault timing and active fault identification in basement terrains with scattered and small sedimentary basin development. In this last chapter we combine the discussed approaches, linking the remote sensing and seismic observation methods to the brittle-tectonic theories and elaborating existing models for structural graben evolution. We discuss the approach of the actual case-study dealing with the structural evolution of Altai-Sayan. As the extensional structures in a region dominated by transpressional deformation are studied, both tectonic regimes, their interactions and relations will be considered here.

5.2 The interpretation of structural elements

Structural elements are features of geologic nature observed on space imagery, on seismic profiles and in the field, which we equate with the classical structures addressed by structural geology (graben, basins, domes, folds, faults, joints, etc.). Fracture and lineament analyses from space imagery often reveal a remarkable persistency of trends over both cratonic areas and mobile belts. Structural elements often display a systematic geometrical relationship with respect to lineaments. These elements can be interpreted in terms of extensional, oblique-slip or contractional tectonics and are related to basin formation-, deformation- and/or uplift histories. The question how features diagnostic for particular geodynamic settings can be recognised by the tools used in the present study is treated. All methods have some limitations: Remote sensing only permits the geometric analysis of structural elements that intersect with the ground surface. The formulated criteria of recognition are suggestions which must be substantiated by field-observations of sedimentological and structural nature, all of utmost value in this matter. High resolution seismic reflection data give only information about the upper few hundred metres of the investigated basin, and extrapolation to depth necessary for the interpretation of the observed structures often remains tentative. The ignorance about the lower crustal- and mantle lithosphere rheology restricts the extend to which we can reliably interpret the observed structures. Nevertheless, inferences for the timing and development of the studied structures can be made from the interpretation of the multi-disciplinary observations.

5.2.1 Extensional tectonics

The most simple (but adequate as a staring point) tectonic models relate extensional tectonics in pure shear to doming and subsequent rifting and report on the classical clay-experiments implemented by H. Cloos in the late thirties [Cloos, 1939]. This set up corresponds to the classical Andersonian classification of planar non-rotational normal shear faults, the axis of maximum compression between which is vertical [Anderson, 1951]. Regional up-arching of the lithosphere produces tensional stresses in the crust. Early experiments show that essential features in extensional tectonics are graben developing perpendicular to the axis of maximal tension and parallel to the axis of maximal horizontal compression [Cloos, 1955]. Graben can either form in regions of crustal extension, or on the crests of broad anticlines related to contractional tectonism (crestal collapse graben). The transition from compressional to tensional stress regimes in the crust is gradational and depends on a combination of factors which are external to the stress system. These factors force the relative position and magnitude of the three principal stress directions to rotate gradually (cf. fig. B.4). The Cloos-experiments show that tensile fractures and shear features form subsequently in the process of doming. The shear fractures are symmetrical around the direction of maximum compression (fig. 5.1a and fig. 5.2). The tension (delational) fractures are parallel to the σ_1 direction. Tensional regimes might consequently be recognised by a combination of compressional shear (mode III normal faults) and tension fractures (mode I tension gashes, dykes, tension joints), as shown in fig. 5.1b. Both conjugated shear planes of the stress system immediately preceding the onset of doming and incipient rifting are sometimes recognised (fig. 5.1a). The bounding rift faults thus show a typical indentation pattern determined by the intersecting shear planes (fig. 5.1b).

During the earlier stages of uplift, the least compressive stress reaches a zero value or becomes tensile while the maximum compressive stress remains positive and larger than the vertical stress (fig. 5.1). S_{Hmax}, at the onset of doming corresponding to σ_1 (fig. 5.1a, gradually decreases to become σ_2 (Fig. 5.1b). This situation degrades progressively with continuing doming until the maximum stress becomes the gravitational vertical direction. After this permutation of stress axes, the morphology of the boundary faults differs from the foregoing one. The originally vertical shear faults tend to rotate about a horizontal axis (tilt) during the horizontal extension. The sinking floor is isolated by vertical and straight or curvilinear faults without any indentation caused by shear fractures which, in this case, intersect along a horizontal axis (fig. 5.1c).

The Cloos-model of tilted blocks in a symmetric graben system is still in use and valid in many instances, and was elaborated further in order to quantify the boundary conditions and stresses. Bott [1981] quantified the graben formation mechanisms for active rifting, where mantle upwelling and subsequent doming precedes graben formation [Bott, 1981]. The latter occurs trough (brittle) wedge subsidence of the upper crust above a weak (visco-elastic) lower crust. Horizontal tensile stresses of >200 MPa are required for a graben of several kilometres deep to develop. Bending stresses related to the up arching alone are insufficient to provide this tension. Assuming a ductile lower crust, the existence of a (topographic) surface load together with a low density mantle wedge however can provide the necessary tensile stress due to vertical buoyancy differences.

Structural considerations complemented the graben formation models by mechanisms explaining the modes of extension and evolution of sedimentary basins by generation of listric rotational and/or planar non-rotational faults [Wernicke & Burchfiel, 1982; Gibbs, 1984; Gibbs, 1990]. These models emphasise the development of asymmetric half-graben. In these instances, the shear stresses which necessarily follow any vertical displacement of a tridimensional body will dissipate by creating listric fault bounded blocks with the geometry of major landslides. The point of departure of this simple-shear scheme is lithospheric stretching along some low-angle detachment, confined between vertical weak zones. Horizontal tension or vertical compression does not automatically bring about the symmetric shear faulting put forward in the Anderson classification and the dependence of the strain on the prevalent stress field, unlike Anderson's model, is undetermined in such cases.

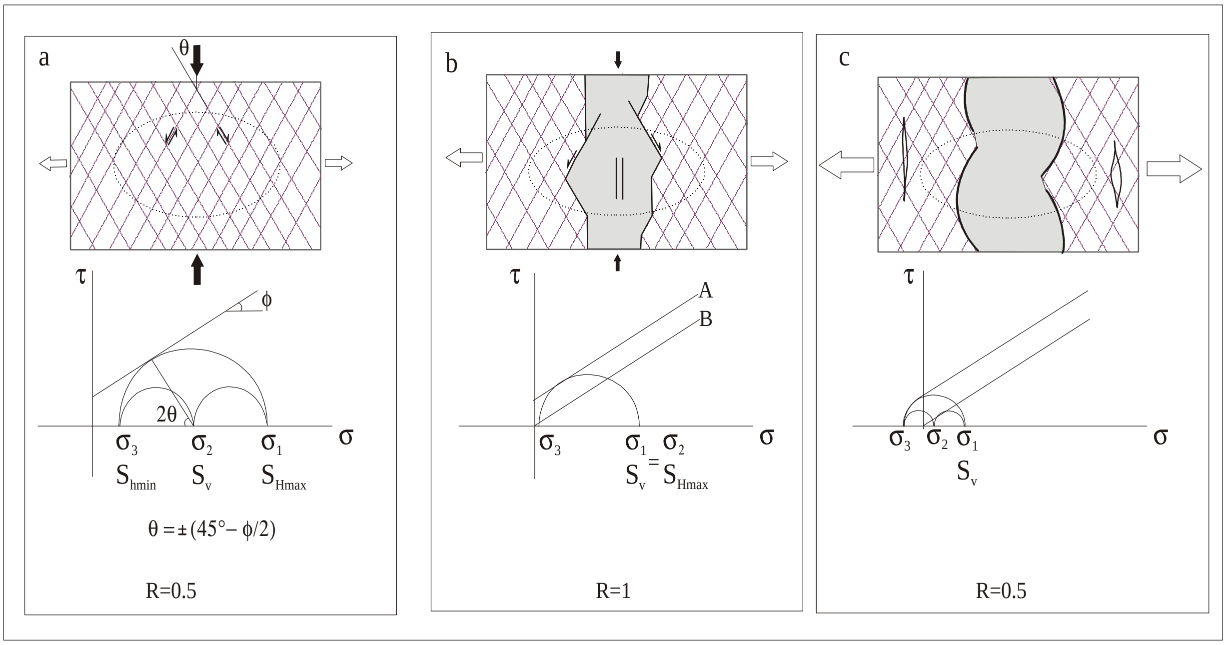

Fig. 5.1 Conceptual cartoon of the evolution of fractures and stress state during the up-arching of a rock body. The ellipse shows the strain directions; a) Strike slip stress regime with conjugated shears faults (stippled lines) at the onset of up-arching. Shear sense is indicated; b) As the stress ellipsoid rotates (σ_1 gets vertical and S_{Hmax} becomes σ_2), extensional shear along the conjugated shear planes (with strength envelope B, [Sibson, 1985; Ranalli and Yin, 1990]) together with neoformed normal shear faults (with strength envelope A) form a graben in a strike-slip-tensional regime. Thick black lines are shear faults; c) Final tensional stress regime. σ_3 equal to the tensile strength causes tensile fractures (black gashes) and normal faulting (black lines).

These models treat all normal faults as one class, unrelated to their driving mechanism. One could, however, distinguish gravity-driven extensional faults induced by slope failure from crustal-stretching faults accommodating the upper crustal part (seismogenic layer) of a whole lithosphere strain [Roberts and Yielding, 1994]. The importance of the distinction is that gravity driven normal faults may be unrelated to lithospheric extension. They generally form in soft sediments, having very little strength, and hence can develop in virtually aseismic regions. The large extensional faults affecting the whole seismogenic layer generally seem to have a planar geometry and dip 30°-60° [Jackson and White, 1989; Kusznir and Ziegler, 1992]. Space problems are solved by assuming plastic flow in the lower crust accommodating the brittle extension within the seismogenic layer [Govers and Wortel, 1995]. These authors argue that most of the low angle normal faults observed in nature are not active, but result from inactive rotation due to subsequent movements on other faults. However, this topic remains debated, for example by Wernicke (1995), who states that geological and geophysical indications for active slip on low angle normal faults do exist. They can be explained by aseismic creep and rotation of the stress trajectories at depth [Wernicke, 1995]. The existence of detachment faults in metamorphic core complexes has been explained by the mature evolution a basic decollement (decoupling zone between crust and mantle) [Brun and Beslier, 1996]. The asymmetric structures in ancient and modern extensional domains, generally attributed to simple shear deformation with fault rotation and detachment development, can also be interpreted as remnants inherited from previous, upper crustal deformation events [Dunbar and Sawyer, 1989]. The latter are controlled by gravity-driven faults (sub-seismogenic layer scale), which are generally listric. In the present study, the gravitational faults are of primary importance in the extensional domains studied.

The mechanics of continental extension were further elaborated in several models, combining McKenzie's pure-shear model [McKenzie, 1978] with Wernicke's [e.g. Wernicke, 1981] simple-shear extension along a detachment surface [e.g. Lister and Davis, 1989; Kusznir and Ziegler, 1992]. The question about rift symmetry related to lithospheric layering and variations in lithospheric strength profiles was investigated trough analogous models [Allemand et al., 1989; Allemand and Brun, 1991]. These models show that a 3 layered lithosphere (upper crust - lower crust - mantle) with a low coupling between the ductile lower crust and the strong (and ductile) mantle produce narrow, symmetrical full graben while a strong coupling induces wide deformation zones with half graben forming simple shear [Allemand et al., 1989]. More recent conceptual models relate the graben asymmetry of most continental rift basins to the locking of one of the initial conjugated normal faults forming the subsiding wedge, when the other fault becomes the leading border fault system of the developing half graben [Scholz and Contreras, 1998]. Finally, it is noteworthy to mention the experimental confirmation to the idea that the main controlling factor in continental graben geometry is the strength ratio between the brittle and the lower crust, highly depending on the extension rate [Michon and Merle, 2000].

Whether the crustal extension, lithospheric thinning and continental breakup is a result of convective asthenospheric upwelling or of horizontal far-field stresses generated by plate interactions remains debated. The tensional stress fields responsible for lithospheric extension can be caused by crustal thickening and consequential gravity instabilities, thermal doming, vertical loading of the hinterland with elastic thinning of the foreland, pure shear tension caused by diverging motion at a continent-anchored trench couple or secondary tension along a load-free edge of a plate under compression [e.g. Ziegler, 1995; Ruppel, 1995].

In situations of vertical tectonics in the Andersonian sense, the azimuth of rifting corresponds to the direction of the main horizontal stress S_{Hmax} after permutation of the principal stress axes. Slip lines are symmetrically distributed around the axis of rifting.

As we already mentioned in the discussion of the Central Asian Deformation Zone in chapter 1, the role of lithospheric anisotropy and crustal heterogeneity on the location and propagation of rifting is important. Active continental extensional tectonics generally is localised along the mobile belts bordering cratonic regions, and structural inheritance influences the rift propagation in many cases. Also continental breakup mostly seems to occur along reactivated mobile belts between cratons. The southern Atlantic ocean for example opened (in the Jurassic) along the Neoproterozoic belt separating the Kalahari and the Rio de la Plata craton, parallel to the tectonic fabric of the basement [Chang et al., 1992]. The mechanism, although, should not be taken for granted, as modelling of vertical weak zones in elastic plates shows that the deviatoric stress field in the weak zones can be higher than the one in the enclosed relatively strong blocks [Govers and Wortel, 1995]. However, many Cenozoic graben faults rejuvenated crustal structures whenever those were properly oriented with respect to the tensional stress. In the Malawi rift, Proterozoic basement structures belonging to the Ubende belt (separating the Tanzania and Zimbabwe cratons) represent the basic anisotropy along which active extension takes place [Ring, 1994]. Also for the Baikal rift, the Proterozoic Primorski shear zone, delimiting the Angara craton, is reactivated as the main active border fault in the present-day rifting [Theunissen et al., 1993; Delvaux et al., 1995; Delvaux et al., 1997; Levi et al., 1997]. Although those structures have to be properly oriented (with a small angle to the regional S_{Hmax} direction), they can influence and guide the local extensional directions. Also the occurrence of transfer faults, accommodating extension by oblique slip along fault segments oriented at high angles to the graben trend can be influenced by pre-existing basement anisotropy [Gibbs, 1984]. Such transfer faults allow (or accommodate) the different slip rate on different normal fault segments and allow the extension to transfer style and activity along the graben. Preferred orientations of olivine in mantle rocks, caused by continental assembly-related deformation during past orogenies, entails a mechanical anisotropy in the deeper parts of the lithosphere. This penetrative fabric reduces the (lithosphere) strength for extension normal to the foliation, favouring rifting parallel to the existing fabric [Vauchez et al., 1997] .

5.2.2 Contractional tectonics

The variety of crustal deformation with a horizontal principal compression direction is wide and of particular interest for the studied region. The analysis of the directions and geometry of the structural elements relies on the fundamental structural associations related to transcurrent tectonics (strike-slip and oblique-slip settings), low angle thrust and high angle reverse fault environments and on the considerations about the stress distribution inside an inhomogeneous segmented crust composed of crustal blocks separated by vertical weak zones. Such considerations are necessary to understand the active tectonic evolution and active fault movements in the Altai-Sayan study area, as a whole dominated by this transpressive deformation. Before discussing the specific properties of oblique tectonics we will consider the end members of the system: thrust tectonics and pure strike-slip transcurrent tectonics.

Thrust tectonics

It is essential while establishing the presence of low angle thrust systems that the main thrusts (floor and roof in a duplex) and related shear zones are recognised. Unless folded and outcropping in windows, these low dipping shear zones are undetectable by means of remote sensing. Other lines of evidence do nevertheless permit to establish the particular tectonic style of low-angle thrust belts, especially the aspects of the oblique-slip deformation along lateral ramps.

Many fold-thrust structures have arc-shaped outlines. Tapponnier & Molnar [1977], a.o., use the festoon-shaped fault trace in low-angle thrusts as a distinctive criterion, in contrast with the straight, linear expressions of normal faults. Remark that repetitions of lobate outlines already served for recognising the margins of some sedimentary basins. This is a complicating factor, however not surprising, because also the vertical sections through sequences of listric normal faults and low-angle thrusts are almost identical [Gibbs, 1984].

Generally, back-thrusts are also curvilinear in plan view. One combination of leading thrust/back-thrust looks on imagery as the eyed texture of "augen gneiss", the frontal ramp of the leading thrust facing one end, the back-thrust facing the other end and closing in on the main thrust. One eye often is fully circumscribed by lineaments. The anticlines on top of both the thrust faults are disposed obliquely and en echelon on the lineament directions, leaving in between a synform on top of the leading thrust sheet.

Characteristic for thrust systems are shear zones parallel to the direction of tectonic transport. The shear zones may be the expression of tear faults, culmination walls and lateral ramps. They are all transverse, perpendicular to oblique, to the fold structures. They are often very prominent on satellite imagery. If the decollement follows paths imposed by basement topography, the tectonic transport and convergence is not necessarily parallel to the shear zones and the moving thrust sheets deform while sliding along. The frontal ramps of successive waves of thrust sheets mostly are arcuate in a horizontal map section. In case tectonic transport is not parallel to lateral ramps, the frontal arcs are oblique to the latter.

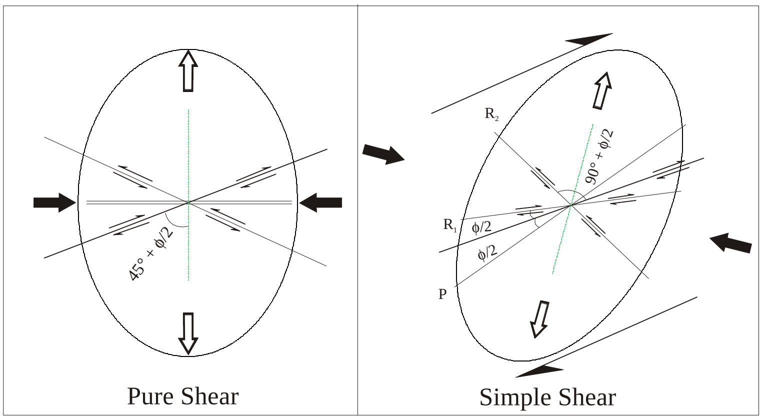

Fig. 5.2 Comparison between pure-shear and simple-shear mechanisms. Coulomb-Anderson model shown on the left side. Riedel model on the right. Black arrows indicate the (incremental strain) shortening axis, white arrows the extensional axis. The long axis of the strain ellipse is at 45° to the moving boundaries (after Sylvester, 1988).

It is often difficult to tell the difference between basins created in strike-slip or thrust settings because the morphology of basins surrounded by the structural highs of active and inactive frontal ramps and laterally confined by ramps or connecting thrust splays, is similar to that of undeformed rhombic pull-apart basins. The general environment of transcurrent regimes, the pattern and the sense of movement (same sense along master faults in vertical simple-shear, opposite sense along lateral ramps in moving thrust sheets) must allow distinction. Windows in duplexes and the combination frontal-lateral ramp may, in some instances, be reminiscent of the typical morphologies of vertical basement tectonics and drape folding. It is however believed that a lineament study may help in discerning between possibilities because vertical basement adjustments should be more intimately linked to long linear basement fractures than thrust tectonics would do.

5.2.3 Transcurrent tectonics

Strike-slip faulting is one of the most important deformation mechanisms in intraplate settings. Terminology and classification was basically ‘standardized’ in the review-article by Sylvester [1988] and will not be discussed here. Our discussion will be limited to the specific mechanics of strike-slip deformation.

Pure shear deformation produces conjugated strike-slip faults, folds and normal faults as shown on figure 5.2. Conjugated fault movements in pure shear deformation can accommodate non-rotational bulk strain as long as both sets operate simultaneously. Because of space problems at large displacements, movements along the faults are limited and remain relatively small. Most large faults culminating major offsets consequently operate through simple shear.

In many continental active zones, convergence is accommodated by distributed simple shear deformation operating through block rotations. The mechanical behaviour of pre-existing structural discontinuities in such settings is supposed to govern the orientation of the active shear zone, which adjusts itself to the movements on reactivated pre-existing fault orientations in order to take up the motion between the bounding blocks or plates [e.g. McKenzie and Jackson, 1983].

The possibilities of block rotations in strike-slip tectonics with continuous (distributed) deformation is of particular interest for the study area. Block rotations, however but by paleomagnetic measurements, are not easily diagnosed. The deformation within a region can be taken within a couple of strike-slip faults. Simple (purely rotational) shear coupling basement blocks deforms overlying cover in a very characteristic way. Secondary folds and faults have an oblique relationship with respect to the moving vertical master faults and related shear zones. They tend to rotate in the direction of the shear couple with ongoing deformation (fig. 5.2 and fig. 5.3). As slip on the rotating faults is bounded by the regional stress orientation, rotation of the faults inside the shear couple is limited (fig. 5.3). Ongoing rotation will induce the formation of new faults with orientations controlled by the Mohr-Coulomb failure criterion [Nur et al., 1986].

Thus, fault rotation is limited by the cohesive rock strength, coefficient of internal friction and principal stress difference (rock strength). For simple shear between two parallel strike-slip faults, tectonic elements can form as shown in figure 5.2. Because the deformation is rotational, the orientation in plan view of the tectonic elements between shear zones varies from highly oblique to near parallel to the shear zone [Thomas, 1974]. The latter situation occurs in advanced stages of simple shear coupling or results from the deformation at low, ductile, crustal levels. Foliation and other structural elements are seen to curve towards the vertical shear zones where they may become folded along vertical axes. En echelon occurrence of the secondary structures (uplifts, drag-folds, faults) can sometimes indicate the shear sense.

Somewhat different are the geomorphologic and tectonic consequences of transcurrent processes with simple shear strain in an anisotropic basement previously segmented by vertical discontinuities. Along strike of the vertical structures, blocks are uplifted (segments sub-parallel to the least compressive stress) or down-faulted (segments sub-parallel to the most compressive stress) from the basement. In cases of sedimentary cover, it follows the relative movement of the basement blocks by drape folding.

The simple shear rotation model clearly is apt to studying ductile shearing strain and deformation of soft layered cover sequences. Whether it is also adapted to clarify the tectonics in brittle basement blocks, what many authors do, is a question we still have to consider. Recent studies analysed block rotation based on fault surface trace analysis and analogue modelling. Souriot and Brun [1992] argued that the Afar triangle in the horn of Africa undergoes dextral rotation due to the existence of a dextral simple shear couple induced by the sinistral rotation of the Danakil block.

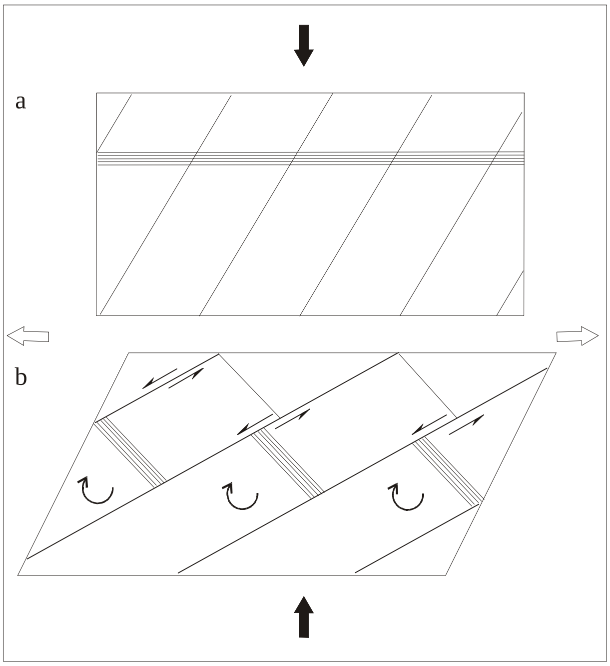

Fig. 5.3 Example of fault- and fault block rotation in strike-slip deformation. Arrows indicate stress axes. a) formation of strike-slip faults according the Coulomb criterion. b) rotation of faults and fault blocks causing normal stress increase and shear stress decrease on the faults, till they become sealed.

Anyway, the simple-shear approach in basement terrains should be used cautiously, and the complication induced by rock heterogeneity and basement anisotropy should be beared in mind.

A typical structural pattern in a region which prevalently underwent simple shear strike-slip tectonics is that of relatively long and straight shear zones delineating corridors in which sub-parallel tectonic elements are arranged in en echelon fashion. The faults proper are long (tens of km), deep and narrow furrows cutting through the landscape as straight, or slightly undulating lines. Riedel shears can develop inside the corridor. They are shorter than the master faults to which their directional attitude is oblique. They mostly do not extend far beyond the moving faults from which they originated. Second order faults together with the inhomogeneous nature of the basement and the cover complicate this simple image at a close view but, at satellite distances, this general statement should hold. Extensional and contractional processes operate contemporaneously. Continuing rotation inside the shear couple will produce a complex basin morphology and deformational history. Extreme development of basin partitioning will thrust one basin against the other.

5.2.4 Oblique-slip tectonics

Strike-slip transcurrent and thrust tectonics are two end members of a system in which rock sequences are transported between two confining surfaces. In pure strike-slip, the sequences are folded along vertical axes between vertical confining surfaces or zones of strike-slip faulting. In case of vertical extension, folding occurs along horizontal axes. The confining surfaces in this case are the floor and roof thrusts of the duplex. In oblique-slip, the vertical and horizontal components combine.

In the latter case, the deformation deviates from the simple shear and pure shear cases discussed before. Strike-slip and extension respect. contraction act simultaneously, causing transpression respect. transtension (fig. 5.4). Transpression and transtension are *strike-slip deformations that deviate from simple shear because of a component of, respectively, shortening or extension orthogonal to the deformation zone* [Dewey et al., 1998]. Transpression or transtension occurs in many tectonic settings, e.g. in undulating strike-slip zones, where respectively shortening and extension (pure shear) and wrench deformation (simple shear) occur simultaneously at respectively restraining and releasing bends. Transpression occurs in zones where the direction of the strike slip fault approaches that of the least compressive stress in the acting system and restraining bends are formed. Transpression and transtension cause the long axis of the horizontal strain ellipse to develop at respectively lower and higher angles to the zone boundaries than in simple shear strike slip (fig. 5.4). For example, this causes in the case of transtension, Riedel shears and tensile fractures to develop at very low angles to the bounding zones (fig. 5.4).

Vertical shear zones, sub-parallel to the orogenic belt, oblique arrangement of the structural grain en echelon with respect to the shear zone, rotational S-shaped rhomb basins and uplifts bounded by strike-slip fault terminations are fundamental structural features of transcurrent tectonics which can be sensed from space. A combination of these observations determines the existence of a prevailing transcurrent setting.

A distinct kind of extensional tectonics is typical in settings dominated by compressive oblique-slip. Fault directions departing from the optimal shear directions in a stress system with vertical intermediate stress and approaching the direction of highest compression might experience tensile stress. This situation often leads to the creation of pull-apart basins. The same basins can also be formed in areas of overlap between the extremities of strike-slip faults. Pull-apart basins are the result of a rotational strain and often display a characteristic S- or Z-shaped sigmoidal geometry. Numerical models of pull-apart basins show the important influence of the upper crust - lower crust and crust-mantle coupling on the graben morphologies. If such coupling is strong, the deformation pattern depends on the width of the shear zone at depth [Katzman et al., 1995]. In this model, wide and shallow basins develop in the upper crust when the underlying shear zone is wide, and vice versa. The symmetry of the graben depends largely on the amount of overlap between the en echelon strike-slip border faults of the pull apart zone. Large overlap between the en echelon border faults produce symmetrical full graben around the centre of the overlap, changing to asymmetric half graben morphologies away from the centre.

Some criteria concerning the geometry and morphology of graben axes and faults surge from the various proposed genetic models. Apart from the indentation of some marginal faults and the linearity of others, the sigmoid prolongation of the far ends of several rifts in down-faulted structures which are oblique to the trend of the main longitudinal rift axis, seems a general characteristic of graben. Many of the, e.g. East African, graben in fact are successions of single depocentres disposed en echelon [Rosendahl et al., 1992; Ebinger et al. 1984]. Each echelon is separated by the mentioned oblique cross-faults (accommodation zones). Typical examples of sigmoid echelons juxtaposed to composite graben are found on images of the Tanganyika [Rosendahl et al., 1992] and the Baikal [Sherman, 1992; San'kov et al., 2000] rifts. The continuation of the border faults at depth, their relation with secondary fractures and the occurrence and architecture of accommodation zones or transfer zones is preferentially studied by interpretation of seismic lines. The occurrence of curvilinear graben border fault segments is used as a distinctive criterion in the recognition of this particular tectonic setting of along axes segmented asymmetric graben [Rosendahl, 1987; Scholz and Contreras, 1998]. The arcs are generated by the listric fault processes and are comparable in shape and origin to the arcs seen in compressive low-angle thrust settings [Gibbs, 1984]. On imagery, the curvilinear border faults or arcs are set inside (and link) the mentioned pre-existing conjugate lineaments (fig. 5.1) [Morley, 1988].

Analogous modelling of rifting at a slow oceanic spreading ridge indicates that oblique extension (not perpendicular to the rift axis) in transtensional settings cause en echelon half graben separated by transfer zones [Dauteuil and Brun, 1996]. This gives yet another possible way for the typical half-graben structures to develop.

The ratio length/width is mostly close to 3 or larger in mature pull-apart basins. Their long axis is sub-parallel to the direction of the overstepping oblique-slip faults between which the basin develops. A marked off-centre position of the depocentres is a critical property in favour of rhombic strike-slip basins [Nilsen and Sylvester, 1995].

Triangular basins might form in the divergent edge between two intersecting strike-slip faults with opposite sense of relative movement. The complicated stress pattern related to these simple-shear settings will result in variable deformation and complicated vertical and rotational movements [e.g. Woodcock and Schubert, 1994].

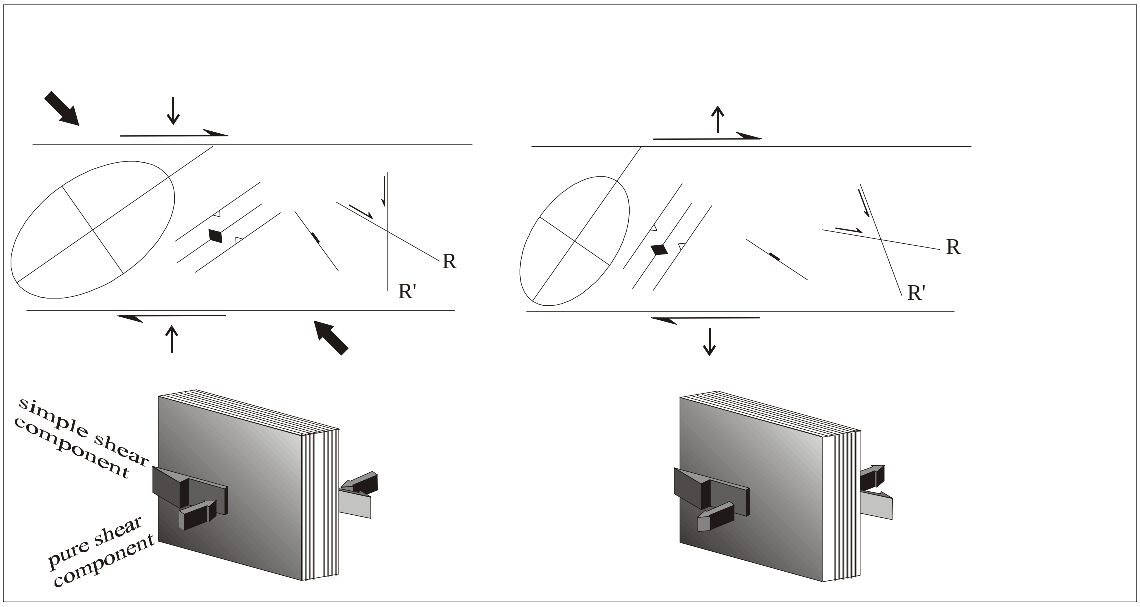

Fig. 5.4 Transpression (left) and transtension (right) in a right-lateral shear zone. Formation of Riedel (R), conjugated Riedel (R’) folds (double black triangles), reverse faults (open triangles) and normal faults (black rectangle) with respect to the shear zone boundaries are indicated. The finite strain ellipse makes a lower, respectively higher angle with the shear zone boundaries in transpression/transtension than in the case of simple shear (fig. 5.2). Modified after Sylvester [1988] and Fossen and Tikoff [1998].

Satellite imagery shows the vertical component of oblique-slip or the total component of the deformation in regions affected solely by strike-slip. The image of the strike-slip corridors (lineaments) between which faulting and folding are confined, is the rotated image of thrust duplexes.

5.3 Timing reconstructions

The problem of the identification of active faults and of timing reconstructions of fault movements in basement terrains is difficult to address, and requires a multi-disciplinary approach, where all possible information about a given fault is combined, leading to most probable interpretations of the observed structures. In the present study, we tried to know as much as possible about the structural setting of the studied faults, onshore and offshore, where remote-sensing, bathymetric- and seismic profiling together with structural fieldwork and gas-geochemical characterisation of the fault zone form the basis of the interpretation. In the second part of the study, we will discuss the application of the multi-disciplinary approach for active fault identification to the Teletsk basin region in Altai, where we were able to combine all aspects of the discussed approach.

5.4 Concluding remarks

The images of structural expression of continental deformation based in the conceptual-, analogue- and numerical models presented in this chapter are basically limited to the Andersonian classification of tectonic regimes with one of the principal stresses vertical. Fortunately, this seems to be the most general case in the earth crust. Unfortunately, the more geologists scrutinise nature, the more cases and situations they find where Andersonian theory is violated [Yin, 1989; Yin and Ranalli, 1992 and references therein]. For example, horizontal shear in the ductile part of the crust causing horizontal shear at the base of the brittle crust necessarily rotates the principal stresses away from Andersonian orientations. While reasoning about the nature and mechanics of the studied deformation, we should remain critical and aware of the limits of our current models, and we should not refrain from questioning the proposed models.

The question about the coupling of deformation between the crust and the mantle is a persisting one, exclusion about which would bring clarity to the mechanisms of crustal deformation and their meaning and position in continental tectonics. Determining whether and how continental crust structures are connected to the (mantle) lithospheric movements at depth is crucial for attributing them to specific continental deformation models. The mode of block rotations in simple shear settings is also related to the occurrence of a detachment zone, without which such rotations are unlikely [Sylvester, 1988; Katzman et al., 1995]. Oblique convergence between continental blocks or plates often is partitioned into pure Andersonian tectonic regimes with horizontal and vertical principal stress (and strain) directions [e.g. Molnar, 1992; Teyssier and Tikoff, 1998]. The reason for such partitioning is to be found in the decoupling of the movements between the (partitioned) crust and the (penetrative deformed) mantle lithosphere, for example along low angle detachment accommodation zones [Teyssier and Tikoff, 1998]. Although oblique slip deformation zones do occur, possibly related to strong coupled systems, the existence of a weak lower crust causes decoupling and induces partitioning of the deformation [Richard and Cobbold, 1989]. Also for the southern Altai, oblique convergence seems to be partitioned in pure strike-slip and thrust faulting [Tapponnier and Molnar, 1979] with the development of transpressional flower structures [Cunningham et al., 1996] and ramp basins [Delvaux et al., 1995, Delvaux et al., submitted]. In this study, we will provide detailed field information about the surface expression of oblique convergence (transpression and transtension) on the northeastern part of the Altai. We will show how movement is partitioned in local and narrow extensional graben, undulating strike-slip systems with restraining, releasing, contractional and extensional zones and local zones of oblique slip. All these settings are expressions of the accommodation of movement to the far field stresses and velocity fields induced on the Central Asian part of the Eurasian continent by various mechanisms, the main of which are continental collision, slab subduction and mantle plume activity. Their likeliness to influence the study region will be discussed in the next part of this study, based on the applied methodology of interpreting the surface expression of the deformation, complemented by available geological and geophysical data.