The Ailao Shan–Red River shear zone (Yunnan, China),
Tertiary transform boundary of Indochina

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Abstract

The Red River Fault zone (RRF) is the major geological discontinuity that separates South China from Indochina. Today it corresponds to a great right-lateral fault, following for over 900 km the edges of four narrow (<20 km wide) high-grade gneiss ranges that together form the Ailao Shan–Red River (ASRR) metamorphic belt: the Day Nui Con Voi in Vietnam, and the Ailao, Diancang and Xuelong Shan in Yunnan. The Ailao Shan, the longest of those ranges, is fringed to the south by a strip of low-grade schists that contain ultramafic bodies. The ASRR belt has thus commonly been viewed as a suture.

A detailed study of the Ailao and Diancang Shan shows that the gneiss cores of the ranges are composed of strongly foliated and lineated mylonitic gneisses. The foliation is usually steep and the lineation nearly horizontal, both being almost parallel to the local trend of the gneissic cores. Numerous shear criteria, including asymmetric tails on porphyroclasts, C–S or C'–S structures, rolling structures, asymmetric foliation boudinage and asymmetric quartz (c) axis fabrics, indicate that the gneisses have undergone intense, progressive left-lateral shear.

P–T studies show that left-lateral strain occurred under amphibolite-facies conditions (3–7 kb and 550–780°C). In both ranges high-temperature shear was coeval with emplacement of leucocratic melts. Such deformed melts yield U/Pb ages between 22.4 and 26.3 Ma in the Ailao Shan and between 22.4 and 24.2 Ma in the Diancang Shan, implying shear in the Lower Miocene. The mylonites in either range rapidly cooled to ≈300°C between 22 and 17 Ma, before the end of left-lateral motion.

The similarity of deformation kinematics, P–T conditions, and crystallization ages in the aligned Ailao and Diancang Shan metamorphic cores, indicate that they represent two segments of the same Tertiary shear zone, the Ailao Shan–Red River (ASRR) shear zone. Our results thus confirm the idea that the ASRR belt was the site of major left-lateral motion, as Indochina was extruded toward the SE as a result of the India–Asia collision. The absence of metamorphic rocks within the 80 km long “Midu gap” between the gneissic cores of the two ranges results from sinistral dismemberment of the shear zone by large-scale boudinage followed by uplift and dextral offset of parts of that zone along the Quaternary Red River Fault. Additional field evidence suggests that the Xuelong Shan in northern Yunnan and the Day Nui Con Voi in Vietnam are the northward and southward extensions, respectively, of the ASRR shear zone, which therefore reaches a length of nearly 1000 km.

Surface balance restoration of amphibolite boudins trails indicates layer parallel extension of more than 800% at places where strain can be measured, suggesting shear strains on the order of 30, compatible with a minimum offset of 300 km.
along the ASRR zone. Various geological markers have been sinistrally offset 500–1150 km by the shear zone. The seafloor-spreading kinematics in the South China Sea are consistent with that sea having formed as a pull apart basin at the southeast end of the ASRR zone, which yields a minimum left-lateral offset of 540 km on that zone. Comparison of Cretaceous magnetic poles for Indochina and South China suggests up to 1200 ± 500 km of left-lateral motion between them. Such concurrent evidence implies a Tertiary finite offset on the order of 700 ± 200 km on the ASRR zone, to which several tens of kilometers of post-Miocene right-lateral offset should probably be added.

These results significantly improve our quantitative understanding of the finite deformation of Asia under the thrust of the Indian collision. While being consistent with a two-stage extrusion model, they demonstrate that the great geological discontinuity that separates Indochina from China results from Cenozoic strike-slip strain rather than more ancient suturing. Furthermore, they suggest that this narrow zone acted like a continental transform plate boundary in the Oligo-Miocene, governing much of the motion and tectonics of adjacent regions. 700 and 200 km of left-lateral offset on the ASRR shear zone and Wang Chao fault zone, respectively, would imply that the extrusion of Indochina alone accounted for 10–25% of the total shortening of the Asian continent. The geological youth and degree of exhumation of the ASRR zone make it a worldwide reference model for large-scale, high-temperature, strike-slip shear in the middle and lower crust. It is fair to say that this zone is to continental strike-slip faults what the Himalayas are to mountain ranges.

1. Introduction

Field studies of Quaternary faults and compilations of earthquake moment tensors help provide a more precise understanding of the active deformation of continents (e.g. Jackson and McKenzie, 1988; Armijo et al., 1989).

In Asia for example, the fact that slip rates on the largest active faults turn out to be on the order of centimeters per year (e.g. Armijo et al., 1986; Kidd and Molnar, 1988; Peltzer et al., 1989) implies that movements along a few narrow zones absorb much of the present-day convergence between India and Eurasia (e.g. Avouac and Tapponnier, 1992, 1993; Avouac et al., 1993). The increasing number of studies of active faults also leads to more quantitative bounds on the relative importance of shortening mechanisms such as strike-slip faulting and overthrusting. The slip rates on faults in western Tibet and the Tien Shan, for instance, suggest that 30–50% of the present-day convergence between India and Siberia is taken up by strike-slip faulting along the northern and southern edges of Tibet (Armijo et al., 1989; Peltzer et al., 1989; Avouac and Tapponnier, 1993).

In general, however, present-day tectonic styles and rates cannot be extrapolated for long into the past. Because the deformation of Asia started with the onset of collision, prior to ~ 50 Ma (e.g. Besse et al., 1984; Patriat and Achache, 1984; Jaeger et al., 1989), instantaneous rates derived from earthquake moment tensors (e.g. Holt et al., 1991) concern at most ~ 2.10-4% of the collision span, while rates deduced from morpho-tectonic studies (e.g. Armijo et al., 1989) characterize only the last few percent of that span. A full, quantitative understanding of the finite strain induced by collision evidently requires analysis of pre-Quaternary movements and deformation.

Although seafloor-spreading reconstruction (e.g. Molnar and Tapponnier, 1975; Patriat and Achache, 1984) and paleomagnetism (e.g. Achache and Courtillot, 1984; Besse et al., 1984; Besse and Courtillot, 1988) provide estimates of the convergence between India and Siberia since collision began (3500–2000 km), regional finite strains and offsets within the collision zone, whose surface area covers about 107 km², are still poorly known. Field studies of the styles, amounts, and ages of Cenozoic brittle and ductile deformation throughout that zone, coupled with more paleomagnetic studies such as those undertaken by various groups in the last decade (Achache et al., 1983; Achache and Courtillot, 1984; Enkin et al., 1992; Chen Yan et al., 1993; Gilder et al., 1993; Huang and Opdyke, 1993; Yang and Besse, 1993), are needed to discriminate between various finite deformation scenarios, among which the large-scale polyphase extrusion model proposed by Tapponnier et al. (1982, 1986) and Peltzer and Tapponnier (1988).

We present here a detailed study of Tertiary deformation and metamorphism along the Red River Fault zone (RRF), in Yunnan province (China), that sets preliminary results discussed by Tapponnier et
Fig. 1. Sketch map of major tectonic features of Southeast Asia after Tapponnier et al. (1986), Peltzer and Tapponnier (1988) and Armijo et al. (1989). Tertiary sea floor (shaded) and spreading axis in South China Sea are from Briais et al. (1993). Thickness of Tertiary sediments from Working Group on Resource Assessment (1991). Box refers to Fig. 2.

al. (1990), Schärer et al. (1990) and Zhong Dalai et al. (1990) on firmer ground. After an outline of the geologic and tectonic histories of regions surrounding this great fault zone, we describe in detail the kinematics, \(P-T\) conditions, and timing of strain events along the Ailao Shan–Red River (ASRR) metamorphic belt. Our observations show, beyond doubt, that large-scale left-lateral shear followed by a reversal to right-lateral occurred along the ASRR zone in the mid-late Cenozoic. This brings support to the basic tenets of the two-phase extrusion model; namely, the early, collision-driven escape of Indochina towards the SE, and the subsequent change to accommodate the present-day escape of Tibet and South China. We examine quantitative constraints on the amount of extrusion and rotation consistent both with the results of our field study and with South China seafloor magnetics (Briais et al., 1993). Finally, we discuss the bearing of such constraints on the mechanical behavior of the continental lithosphere.

2. Outline of the geology and tectonic history of Yunnan

Yunnan is the southernmost large province of China. Morphologically, it is a mountainous region that blends with the Tibetan highlands towards the northwest, where the elevation of many ranges exceeds 4000 m. Towards the east it forms a broad plateau, about 1800 m high, with more gentle relief, that resembles that of Guizhou province. Geologically, Yunnan straddles three of the main domains that Chinese geologists have long distinguished in this part of Asia: the Yangzi paraplatform to the northeast, the Sichuan–Eastern Tibet (Songpan Ganzi, Sanjiang) fold belts, to the northwest, and the Indochina “block” or Sundaland to the south (e.g. Huang, 1960; Wang Hongzhen et al., 1985). The Red River Fault zone (RRF) separates the later block from the two previous regions, which belong to South China.

This zone has long been recognized as one of the main geological discontinuities of Southeast Asia (Fig. 1) (e.g. Huang, 1960; Huang, 1978; Hutchinson, 1982; Wang Hongzhen et al., 1985). It follows the Ailao Shan–Red River (ASRR) metamorphic belt, that comprises four principal ranges, or massifs, in which mid-crustal rocks have been unroofed: the Ailao Shan, Diancang Shan, and Xuelong Shan, in Yunnan, and the Day Nui Con Voi in Vietnam (Fig. 2). The active Red River and Qiaohou faults (Fig. 5) also follow this zone, which extends more than 1000 km, from the Gulf of Tonghzing to the Hengduan mountains (Fig. 1). Farther north, towards Tibet, the NW–SE-striking RRF zone bends clockwise, exposing the more northerly trend of the Sanjiang (“Three Rivers”) mountain ranges (Fig. 2).

Any assessment of the nature and role of a zone of that length requires knowledge of the geological history of the regions that it crosses. In particular, it is important to understand how the zone relates to major regional tectonic episodes from the Precambrian to the present, and what similarities and differences in paleogeography or strain styles exist on either side of it along strike. Hence, before focussing on strain and metamorphism along that zone we broadly discuss the large-scale geology and tectonics of regions located north and south of it.

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Fig. 2. Structural sketch map of Yunnan province and neighboring areas (box on Fig. 1), resulting from compilation of following maps: for Yunnan area, Geological Bureau of Yunnan Province (1979), Bureau of Geology and Mineral Resources of Yunnan (1983), Bureau of Geology and Mineral Resources of Yunnan Province (1990); for Tibet, Tang Yaoqing et al. (1988); and for surrounding regions, General Geological Department of the Democratic Republic of Vietnam (1973), Earth Sciences Research Division (1977).

X.L.S. = Xuelong Shan; D.C.S. = Diancang Shan. Cities: B.Y. = Bao Yen (Vietnam); Ch. = Chengdu (Sichuan); C.M. = Chiang Mai (Thailand); Chu. = Chuxiong (Yunnan); C.R. = Chiang Rai (Thailand); D. = Deqen (Yunnan); D.B.P. = Dien Bien Phu (Vietnam); H. = Hanoi (Vietnam); J. = Jinghong (Yunnan); K. = Kunming (Yunnan); L. = Lijiang (Yunnan); L.e. = Leshan (Sichuan); Li. = Litang (Sichuan); M. = Mandalay (Burma); Ma. = Markam (Tibet); My. = Myitkyina (Burma); Q. = Qamdo (Tibet); S. = Simao (Yunnan); T. = Tengchong (Yunnan); V. = Vientiane (Laos); X. = Xiaguan (Yunnan); Y. = Yuanmou (Yunnan). Boxes A, B and C indicate areas of Figs. 3a,b and 6, respectively.
Neogene and Quaternary
Jurassic to Eocene
Intrusive rocks
Altai Shan - Red River Metamorphic belt (HT - LP metamorphism)
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2.1. South China block: Yangzi paraplatform and southern end of the Sanjiang belt

2.1.1. Yangzi paraplatform

The Yangzi paraplatform corresponds to the area north of the Red River Fault zone and east of the Jianchuan fault (Figs. 2 and 3). High-grade metamorphic rocks, whose protolith might include the basement of the Yangzi platform, are exposed only along the Yuanmou and Red River faults (Figs. 2 and 3) (Lu Liangzhao, 1989). Elsewhere, sediments ranging in age from mid-upper Proterozoic (Sinian) to Eocene, up to 10 km thick (Wang Hongzhen et al., 1985), cover the basement. Proterozoic and Palaeozoic rocks crop out in two main areas, the Kunming high and the Lijiang mountains that flank the Chuxiong Mesozoic basin to the east and west, respectively (Figs. 2 and 3).

2.1.1.1. Kunming high and Chuxiong basin. South of Kunming, the lowermost sediments are folded, schistosed Proterozoic shales, carbonates and volcanoclastics. They are unconformably covered at places by Lower Devonian conglomerates and sandstones, followed by the Upper Devonian, Carboniferous and Permian shallow-water carbonates that give rise to the south China tower karst (Geological Bureau of Yunnan Province, 1979; Karst Institute of Guilin, 1981; Bureau of Geology and Mineral Resources of Yunnan, 1983). The unconformity testifies to the proximity of the “Caledonian” belt of south China, whose existence is well established in Guangxi province where metamorphic Siluro-Ordovician schists are capped by old red sandstones (Karst Institute of Guilin, 1981). Near Kunming, basalt flows coeval with the Emei Shan basalts of Sichuan are interbedded with the Upper Permian limestones. The basalts are particularly abundant west of Kunming where they include thick pillow-lava units (e.g. north of Midu). Such volcanism indicates regional extension and localized rifting of the carbonate platform at the end of the Permian (e.g. Wu Genyao, 1993).

Mesozoic deposits cover much of the platform in Yunnan. Southeast of Kunming, Lower–Middle Triassic, mostly shallow-marine carbonates are generally conformable on the Permian (Bureau of Geology and Mineral Resources of Yunnan, 1983). West of Kunming, in the Chuxiong basin (Fig. 2), Upper Triassic gray sandstones and marls grade upwards into Jurassic and Cretaceous red sandstones. The lower red sandstones display decametric cross-bedding. Near Lufeng (Fig. 3), they contain fossil Lower Jurassic dinosaurs including the prosauropods *Lufengosaurus* and *Yunnannanosaurus* (Sun and Cui, 1986). The youngest red beds, north of Chuxiong, are of Eocene age (Fig. 3) (Bureau of Geology and Mineral Resources of Yunnan, 1983). To the west, the maximum thickness of the Chuxiong Mesozoic–Tertiary continental sequence is about 6 km (Wang Hongzhen et al., 1985). In spite of the existence of conglomeratic horizons, there is no clear evidence of an important regional unconformity within that sequence, whether in the field or on published maps (e.g. Bureau of Geology and Mineral Resources of Yunnan, 1983). On the other hand, the Upper Triassic, the Jurassic and even the Eocene red beds are unconformable, often at a low angle, upon the Palaeozoic and Proterozoic rocks of the Kunming region, which appears to have formed a broad N–S-trending high or horst since the lower Mesozoic. We conclude that, in Yunnan, the Yangzi platform remained essentially stable from the Upper Triassic to the lower Tertiary. Around Kunming, the basal unconformity of the Mesozoic–Tertiary continental sequence seems to be a distal onlap over a region mostly structured by late Palaeozoic extensional block-faulting.

Cenozoic deformation has been strong. Regional, mostly NE–ENE-oriented shortening has folded the red beds over much of northern Yunnan after the Paleocene (Bureau of Geology and Mineral Resources of Yunnan, 1983; Tapponnier et al., 1986, 1990). Folds with NW–NNW-trending axes, clear on geological maps and satellite images, are well developed in the Chuxiong basin (Fig. 3). In the western part of this basin, the NE-facing limbs of anticlines are often vertical or inverted, suggesting NE-verging thrust décollements at depth (section Ib, Fig. 4). On retrodeformable sections drawn from the 1:500,000 map (Geological Bureau of Yunnan Province, 1979) and field observations, the corresponding shortening is moderate, on the order of 20% (e.g. 20–25 km along section I between Kunming and Midu; Fig. 4). The Tertiary fold axes bend counterclockwise to more westerly trends as they near the NW–SE-striking Chuxiong–Nanhua fault (Tapponnier et al.,
Fig. 3. Structural map of central and southern Yunnan province (boxes A and B on Fig. 2) compiled mostly from Geological Bureau of Yunnan Province (1979), Bureau of Geology and Mineral Resources of Yunnan (1983), Bureau of Geology and Mineral Resources of Yunnan Province (1990), and complemented by LANDSAT satellite image analysis and fieldwork. To discriminate Himalayan structures from Indosinian ones, only folds affecting terranes younger than Upper Triassic are mapped. Diverging arrows indicate anticlines while synclines are indicated by converging ones. Boxes a and b refer to maps of Figs. 9 and 19, respectively. Lines A and D correspond to cross-sections of Ailao Shan shown on Fig. 10. Lines Ia, Ib, II and III are cross-sections of Fig. 4. M corresponds to cross-section of Fig. 28. S refers to Jianchuan–Madeng traverse (Fig. 7a).
Fig. 4. Cross-sections of sedimentary cover in Chuxiong (section Ia, Ib) and Simao (sections II and III) Mesozoic basins. For locations of sections see Fig. 3. On each section shortening of the Mesozoic cover has been estimated between points A and B. Thin, straight, parallel lines represent cleavage.
1990), indicating drag by left-lateral strike-slip along that fault (Fig. 3). A steep, NW–NNW-striking, axial plane cleavage is often observed in the core of the anticlines (section 1b, Fig. 4). Locally, cleavage planes in flattened red shales bear a down-dip linearization, marked by green, slightly elongate reduction spots. The average elongation ratios of the spots, measured at one site west of Chuxiong, are 1.46 in the X/Y cleavage plane and 2.41 in the XZ plane ($X > Y > Z$ being the strain ellipsoid axes). Approaching the Red River Fault zone from the north, the cleavage becomes stronger. North of Gasa, it grades from fracture to intense flow cleavage as the Mesozoic rocks become epimetamorphic slates. The cleavage, which is nearly vertical, bends counterclockwise from a N170°E average strike to N150°E, parallel to the Red River Fault zone and to the foliation in the Ailao Shan gneisses, consistent with left-lateral movement on that zone (Tapponnier et al., 1990). At one site, about 15 km northwest of Gasa and 2 km northeast of the Red River fault, two steep superimposed cleavages may be observed in purplish, probably Triassic, slates and conglomerates. The first, nearly bedding-parallel cleavage ($S_{0+1}$) strikes N140°E, parallel to the Red River fault and bears a clear down-dip stretching lineation. The second cleavage ($S_2$) strikes N160 °E, and defines sigmoidal lenses between the $S_{0+1}$ planes, consistent with left-lateral shear on $S_{0+1}$. These microstructures, which give the slates the appearance of C–S tectonites, seem to result from folding and flattening due to ENE shortening, giving rise to $S_{0+1}$, followed by left-lateral shear, also due to ENE shortening, giving rise to $S_2$.

Folds with NE–ENE-trending axes, almost perpendicular to the predominant folds in the Chuxiong basin, also exist, mostly adjacent to the southern splays of the Xianshuie fault (Xiaojiang, Anning He and Yuanmou faults; Figs. 2 and 5). North of Kunming, such folds affect concordant Jurassic and Permo-Carboniferous beds. South of Kunming, steep, E–W- to NE–SW-striking cleavage is found locally in the Mesozoic sandstones. Northwest of Yuanyang (Fig. 3) along the Red River, E–W cleavage in Paleozoic rocks is parallel to folded, steeply dipping Neogene conglomerates. We interpret these NE–SW to E–W folds and cleavage to result from late Neogene–Quaternary NNW–SSE shortening, compatible with left-lateral slip on the Xiaojiang fault, right-lateral slip on the RRF (e.g. Tapponnier and Molnar, 1977; Allen et al., 1984) (Fig. 5), and with the SSE-directed impingement of the Kunming–Chuxiong block against the northeastern edge of Indochina (the Red River Fault).

This region is presently undergoing active deformation, as attested by the fairly high level of historical and current seismicity (Institute of Geophysics, 1976). Extant fault-plane solutions show that, as in the late Neogene, this deformation is consistent with roughly NNW shortening, but with roughly ENE extension as well (Fig. 5). Right-lateral slip occurred on the NW–SE-striking Quijiang fault during the Tunghai earthquake ($M = 7.7, 4/1/1970$) (e.g. Tapponnier and Molnar, 1977) (Fig. 5a). The focal mechanisms of the 1966 earthquakes on the N–S-striking Xiaojiang fault (Fig. 5) imply left-lateral slip along it (e.g. Tapponnier and Molnar, 1977). A normal component of slip on the roughly N–S faults south of Kunming has created several Quaternary half-grabens, some of them filled by lakes (Fig. 5) (e.g. Tapponnier and Molnar, 1977; Allen et al., 1984).

2.1.1.2. Lijiang mountains. The Lijiang region, between the Cheng Hai and Jianchuan faults (Figs. 3 and 6), is characterized by a thick sequence of Siluro-Ordovician–Triassic sedimentary rocks, mostly clastics and shallow-water carbonates with some interbedded Upper Permian basalts (Bureau of Geology and Mineral Resources of Yunnan, 1983). The lowermost part of the sequence is well exposed north of Lijiang, in the core of a N–S-trending antiform that rises to over 5000 m in the Yulong (5596 m) and Haba Xue (5396 m) Shan (Fig. 6).
There, along the narrow Tiger leap-gorge (Hutiao Xia), the Jinsha river (upper Yangzi), whose course must be antecedent to the growth of the antiform, has cut a ~3500 m deep "cluse" into the folded sedimentary sequence. Micaschists, derived from laminated, crossbedded sandstones and volcanics (Siluro-Ordovician—Lower Devonian?), crop out in the deepest part of the antiform core. They are overlain by white, phlogopite-bearing marbles (Upper Devonian?), in turn overlain by a series of black schists, light-gray calcschists and marbles containing horizons of greenschists and boudins of metabasalts and amphibolites (Carboniferous-Permian?). Most of the marbles and schists are sheared parallel to bedding, the bedding-parallel foliation being folded by the antiform (Lacassin et al., 1993a, 1995). A prominent stretching lineation, folded with the foliation, may be restored to a mean N40°E direction after unfolding about the antiform axis (N175°E). Shear indicators in either limb of the antiform indicate a top-to-the-SW sense of shear prior to folding. The phlogopite-bearing marbles yield a Rb/Sr age of 35.9 ± 0.3 Ma (2σ, isochron defined by several fractions of phlogopite and whole rock, initial 87Sr/86Sr ratio of 0.712) (Lacassin et al., 1995). This late Eocene age probably dates SSW-directed transport (in the present geographical coordinates) on an originally flat décollement, in all likelihood a result of the penetration of India into Asia (Lacassin et al., 1993a, 1995). The well-preserved antiformal shape and the elevation (∼5600 m) of the Yulong mountain, which still stands high above the surrounding country, implies that the folding of that décollement results from the same ∼ENE directed Tertiary shortening that folded the red beds of the Chuxiong basin (Fig. 3). Since transport on the décollement predates the antiformal folding, ENE shortening must postdate 36 Ma in northeast Yunnan.

North of the Diancang Shan, most of the NNE-striking faults that dominate the crustal fabric appear to be presently active (Figs. 5 and 6), as attested by the sharpness of their traces on satellite imagery and in the field and by the seismicity, which includes at least one great historical event (Yongsheng, 1515) (e.g. Institute of Geophysics, 1976; Allen et al., 1984; Wu and Wang, 1988). The faults have oblique, normal left-lateral slip (Tapponnier and Molnar, 1977; Allen et al., 1984) (Fig. 5a). Such movement probably results from bookshelf faulting due to present NW–SE right-lateral shear parallel to the Red River Fault (Fig. 5b). While ongoing movement on those faults is consistent with the currently ∼E–W regional extension, several of them appear to be former W-dipping thrusts related to the ENE, mid-Tertiary shortening phase. The most prominent of these former thrusts, the Chenghai Fault, places Permian volcanics on top of Jurassic sandstones (Bureau of Geology and Mineral Resources of Yunnan, 1983) (Figs. 2, 3 and 6). Inversion of these faults may account for their present-day listric geometry (Wu and Wang, 1988).

In spite of the apparent complexity of the Cenozoic deformation, however, and of stress changes in part related to large-scale clockwise rotations, Tertiary folding of the Yangzi platform appears to result mostly from two successive, nearly orthogonal phases of shortening, oriented ENE and NNW in the present-day geographic coordinates. These phases postdate deposition of the Paleocene red beds and of the Neogene conglomerates, respectively.

2.1.2. Southeastern Sanjiang and the Benzilan–Jinsha suture

West of the Jianchuan fault, northwestern Yunnan lies in the southern termination of the Sanjiang ("Three Rivers") mountain belts of western Sichuan and southeastern Tibet. The belts, which strike roughly N–S, channel the flow of the Jinsha, Lancang and Nu Jiang (Yangzi, Mekong, and Salween rivers, respectively) (Figs. 2 and 6). The high, rugged relief indicates that the Cenozoic deformation of this area, which lies only ∼500 km from the eastern end...
of the Himalayan range (Fig. 1), has been particularly strong (e.g. Wang and Chu, 1988). The regional crustal fabric is characterized by NNW–SSE- to N–S-striking faults bounding long and narrow tectonic stripes that involve mostly Triassic or late Paleozoic rocks (Figs. 2 and 6) (Bureau of Geology and Mineral Resources of Yunnan, 1983). Southwest of the Jinsha river, these roughly parallel fault-stripes bend or abut against the NW–SE Red River Fault zone, whose extension is marked here by the active Qiaohou fault (Allen et al., 1984). This latter fault may be followed northward from the northern tip of Diancang Shan to Weixi (Fig. 6), and farther to Deqen, near the border between Yunnan and Tibet (Bureau of Geology and Mineral Resources of Yunnan, 1983) (Figs. 2 and 6). The N–S fault-stripes are cut at places by shorter NW–SE-striking faults. One such fault (the Zhongdian fault) extends from Deqen to the southeast of Zhongdian (Figs. 5 and 6). The city of Zhongdian is located in a Quaternary basin on a right step along that fault (Figs. 5 and 6) implying that the fault may play a role analogous to that of the Poqu fault, as part of a right-lateral en échelon array connecting the Karakorum–Jiali fault zone of southern Tibet (K.J.F.Z., Fig. 1) to the Red River Fault (Fig. 1) (Armijo et al., 1989).

Three roughly E–W traverses (Jianchuan-Lanping, Judian-Weixi, and Benzilan-Deqen sections; S, T, and U on Fig. 6, respectively), described below, throw light on the stratigraphy and structure of the tectonic stripes located between the Jinsha and Lancang rivers.

The Jianchuan-Lanping section (S on Figs. 3, 6 and 7a) starts in a ≈ 2000 m thick series of continental molasse of mid-Upper Eocene age, overlying Lower Eocene red beds (Bureau of Geology and Mineral Resources of Yunnan Province, 1990). The molasse, composed of massive polymict conglomerates with well rounded cobbles (limestones, red sandstones, volcanics, etc.), is intruded by alkaline, subvolcanic monzonites (Fig. 6) (Bureau of Geology and Mineral Resources of Yunnan, 1983). Titanite and zircon fractions in one intrusion yield an U/Pb crystallisation age of 35 ± 0.1 Ma (Schärer et al., 1994). The conglomerates, which have been uplifted by the W-dipping normal fault that bounds them to the west (Figs. 3, 6 and 7a), form much of the high mountain west of Jianchuan (Laojun Shan, 4247 m). They are only gently warped and rest unconformably on all other rocks, including metamorphic Paleozoic schists to the northeast. To a degree, they resemble the roughly coeval Kailas–Hemis–Noijinkangsang conglomerates of the Indus and Zangbo valleys in southern Tibet. They are the oldest, thickest coarse Cenozoic conglomerates we have seen in Yunnan. Elsewhere, similar rocks are of Neogene age. Important tectonic movement thus occurred prior to the Upper Eocene in this part of Yunnan.

The section continues westwards across folded, mostly W-dipping Triassic rocks. The rocks include green-purple pelites and sandstones with interbedded horizons of green dacitic and pink rhyolitic flows and tuffs, overlain by thinly bedded, fossiliferous (*Halobia* sp.) limestone breccias and black shales, followed by more massive greywackes, dark gray limestones and finally epimetamorphic calcschists. The characteristic association of purple Triassic pelites and green dacites observed here is also found near Luchun, south of the southeasternmost part of the Ailao Shan (Fig. 3). On the whole, this Triassic sequence also resembles that seen in northern Thailand. The rocks apparently belong to the Weixi group (Jen Chi Shuen and Chu Ching Chuan, 1966). That group, of Lower–Middle Triassic age (up to Carnian), is chiefly composed of folded and slightly metamorphic shales and calcschists capped by limestones and sandstones of Norian age. Where well developed, the cleavage in the Triassic rocks strikes ≈ N–S, consistent with roughly E–W shortening. This cleavage becomes particularly strong in the steep, SSE-striking epimetamorphic calcschists just east of the active Qiaohou fault. Close to the fault, the cleavage is folded and affected by numerous kink bands. By contrast, although they dip steeply, the greenish, fossil-plant bearing sandstones and lignite beds that fill the small Neogene Madeng basin just west of that fault (Figs. 3 and 7a) show no cleavage at all. West of this Neogene basin, the section ends near Lanping in the tightly folded, Jurassic–Cretaceous to Lower Eocene limestones and red beds of the Yangbi basin. The NNW-striking fold crests that form the summits of the Xuebang Shan (4295 m) west of the Madeng basin (Fig. 6) imply that these rocks have suffered strong Tertiary shortening in an ENE direction, in present-day geographic coordinates.
The Judian–Weixi traverse (T on Figs. 6 and 7b) first crosses several kilometers of a monotonous sequence of gray, quartz-rich schists (Cambro-Ordovician–Devonian; Bureau of Geology and Mineral Resources of Yunnan, 1983). The schistosity strikes N–S and dips 20–50°W in general. The schists include pyrite nodules, boudinaged quartz veins and a few tens of meters of fine-grained, leucocratic gneisses (probably deformed acidic-volcanic tuffs) with SSW-plunging lineation, and top-to-the-SSW shear senses, as in the Yulong Shan décollement. Westwards, the schists grade into darker carboniferous slates (Bureau of Geology and Mineral Resources of Yunnan, 1983), bounded to the west by a prominent, = N10°E-striking, W-dipping fault (Fig. 7b). The fault has a sharp morphological trace and dams east flowing rivers, creating a series of small Quaternary basins clear on LANDSAT images and air photographs. We interpret this fault to be a Quaternary, apparently active, normal fault (Fig. 5), northward extension of the fault bounding the Eocene molasse farther south (Figs. 6 and 7). Folded, steeply N-dipping, N75°E-striking greenschists, calcschists and thinly bedded limestones, crop out west of the fault. These rocks, which may be of Permian or Lower Triassic age, are intruded by an undeformed,
coarse-grained granite. The western rim of this granite intrudes steeply dipping, massive volcanic rocks including both mafic (dolerites, spilites, and vertical, flattened pillows that strike N110°E) and acidic terms (rhyolites). These bimodal volcanics occur in a sedimentary sequence that comprises folded limestones, red sandstones and purplish pelites (Bureau of Geology and Mineral Resources of Yunnan, 1983). Like that on the Jianchuan–Lanping traverse, this sequence probably belongs to the Triassic Weixi group. The section ends in the Neogene Weixi basin, whose western limit is an E-dipping Quaternary normal fault, bounding the metamorphic core of the Xuelong Shan. The structure and deformation of that range, northernmost gneiss massif along the Red River Fault zone, will be described later.

Between the Mekong and Deqen, section U2 (Figs. 6 and 8) crosses a sequence reportedly Permo-Triassic in age (Bureau of Geology and Mineral Resources of Yunnan, 1983) that comprises flyschoid schists with volcanic rocks of andesitic affinity (including agglomerates and breccias) and typical flyschs. The rocks display steep cleavage and are cut by brittle fault zones. Near Deqen, a nearly vertical N–S-striking fault zone juxtaposes kilometric slices of schists, calcshists, deformed volcanic rocks, micaschists and gneisses, red-purple pelites, and granodiorite, separated by gouge and cataclasite (Fig. 8b). The fault zone, which is a few kilometers across, forms the western boundary of a 5–10 km wide granodiorite massif (Baimaxue Shan, 5137 m; Figs. 6 and 8). Farther east, flyschoid rocks and thick volcanics including pillow lavas are overlain by Triassic limestones. The whole assemblage is strongly folded and unconformably capped by red conglomerates and sandstones of probable Eocene age (Fig. 8b) (Bureau of Geology and Mineral Resources of Yunnan, 1983).

Still farther east (section U1 on Figs. 6 and 8), near Benzilan, an outstanding ophiolitic melange is exposed in a N–S-striking, 3–4 km wide zone. The overall dip of the zone is to the W. From E to W (i.e. from bottom to top) one finds deformed gabbros, a tectonic melange of mafic volcanics and limestone slices, red or purple cherts, serpentinitized harzburgites, more limestones alternating with mafic volcanics including pillows encrusted by cherts, and finally flyschoid sandstones (Fig. 8a). The rocks form slivers separated by cataclasitic zones that locally show normal/right-lateral senses of shear, particularly in the serpentinitized ultramafics. East of Benzilan, a strongly deformed, steep sequence of marbles and calcshists with interbedded horizons of flattened pillow basalts, greenschists and acidic tuffs, probably Permian in age, is exposed along the Jinsha river. The rocks display bedding-parallel foliation, with a = N–S strike, dips ranging between 60 and 80°W, and a strong down-dip stretching lineation. West side-up shear senses imply that this zone was a major ductile E-directed thrust (Fig. 8a).

The Benzilan–Deqen section thus clearly cuts a major suture zone. To the east, one finds the deformed passive margin of the Yangzi platform with the characteristic pillow basalts associated with Permian extensional faulting. To the west, one finds an active margin, complete with the effusive and intrusive rocks characteristic of a calcalkaline volcanic arc and the flyschoid sediments apparently deposited in the back and fore-arc basins. Dismembered slices of a complete ophiolitic sequence attest to the former presence of deep sea floor. The eastward vergence of the thrust affecting the cover of the Yangzi platform is consistent with collision following W-directed subduction under the arc (in the present geographic coordinates), in keeping with paleogeographic evidence (e.g. Sengör, 1984). Final suturing apparently occurred in the Lower Triassic (Liu Baotian et al., 1983). Subsequent shortening and shearing of the suture zone, likely in the Tertiary, could have formed the N–S-striking dextral cataclastic zones along it. The Benzilan suture zone appears to represent the southernmost extension of the Jinsha suture, which separates the Qiangtang block from the Song Pan terrane in western Sichuan and Tibet (e.g. Chang Chenfa et al., 1986; Mattauer et al., 1992) (Fig. 1).

Between Benzilan and Deqen (Fig. 2), small folds and thrusts cut a mid-Late Pleistocene colluvial apron. The thrusts dip 25–50° to the northwest and strike between 40 and 90°E, consistent with NNW, late Quaternary shortening, as observed elsewhere in Yunnan (Fig. 5).

2.2. Indochina and the “Indosinian” sutures

Blocks and terranes located south of the Red River Fault zone form the Indochina “Block” or
Sundaland (Fig. 1) (Hutchinson, 1982). As on the western Yangzi platform, Mesozoic continental red beds cover older sediments and basement over wide areas. West of the Mekong river (Lancang Jiang) a large granite batholith (the Lincang batholith), dated at 270 ± 59 Ma (Rb/Sr on whole rock; Liu Changshi et al., 1989) intrudes a N–S-striking belt of micaschists, metavolcanics and gneisses, about 60 km wide and 400 km long (Figs. 2 and 3). The rocks are interpreted to represent the northern extension, in Yunnan, of the metamorphic and granitic belt of northwestern Thailand (Figs. 1 and 2) (Academy of Geological Sciences of China, 1975; Department of Mineral Resources, 1982; Hutchinson, 1983). In Thailand, the granites are ≈ 200–240 Ma (Rb/Sr on whole rocks; Beckinsale et al., 1979). North of the Simao basin, micaschists and quartzites, intruded by a granite, form the core of the Wuliang Shan dome, northernmost prong of the Lincang belt (Fig. 3). Permo-Carboniferous carbonates, and Lower Triassic limestones and volcaniclastics crop out south of the Ailao Shan, in the core of anticlines near Pu Erh, and along the Mekong valley (Fig. 3) (Bureau of Geology and Mineral Resources of Yunnan, 1983). Undeformed pillow-lavas, of probable Permian age, are well exposed near Mojiang.

The Mesozoic red beds of the Simao basin, whose ages range from Upper Triassic to Eocene, reach a
thickness of 8 km (Wang Hongzhen et al., 1985). They are affected by folds and thrusts that trend NNW–SSE (Figs. 3 and 4). The thrusts are particularly well developed in the eastern part of the basin, between Simao and the Ailao Shan (Fig. 3). The folds appear to be ramp anticlines on generally W-directed thrusts (Fig. 4) and usually correspond to topographic highs, which suggests fairly recent fold growth. NW–SE- to E–W-striking strike-slip faults branch on the thrusts, forming lateral ramps. The fact that anticline cores expose Jurassic red beds near Simao and Permo-Carboniferous limestones more to the east (Fig. 3), suggests progressive eastward deepening of a regional décollement. Near Simao, red Upper Cretaceous–Paleocene shales lie in the hinges of synclines (Bureau of Geology and Mineral Resources of Yunnan, 1983). From first-order retrodeformation of sections II and III, shortening and W-directed transport of the Mesozoic cover appear to be on the order of ≈ 20 km (24%) and ≈ 40 km (33%), respectively (Fig. 4). In the western part of the Simao basin, the vergence of folds and thrusts is opposite, toward the east. At the southern boundary of the basin, near Jinghong, the Lincang batholith is offset at least 40 km by NW–SE left-lateral faults (Bureau of Geology and Mineral Resources of Yunnan, 1983).

Ultramafic bodies and rocks with ophiolitic affinities are exposed along several narrow belts within the Indochina “block”. They have generally been interpreted to mark suture zones (e.g. Hutchinson, 1975; Sengör, 1984, 1987; Hutchinson, 1989), the Song Ma zone, in Vietnam, and the Uttaradit zone, in Thailand, being the most prominent (Fig. 1).

In north Vietnam, about 120 km south of the Red River, serpentinized gabbro-dolerites, harzburgites and dunites from which chromite is mined, mark the NW–SE-striking Song Ma suture zone for at least 170 km (Figs. 1 and 2). The ultramafic rocks, which appear to have been thrust northwards, follow a belt of metamorphic, reportedly Paleozoic rocks. South of that belt, reportedly Siluro-Devonian flyschoid shales and acidic volcanic tuffs are intruded by granodiorites and granites. The age of the Song Ma suture zone is unknown, even though the paleogeography may be taken to imply SW-directed subduction (in the present geographic coordinates) possibly in the mid-Paleozoic. The existence of a regional Upper Carboniferous unconformity (Deprat, 1914) might suggest an upper Paleozoic collision of this active margin (Khorat–Kontum block) with the South China block (Fontaine and Workman, 1978; Helmcke, 1985; Sengör et al., 1988). North of the Song Ma suture, however, in the Song Da belt, Permian basic volcanics, Triassic shales and marine limestones, as well as Cretaceous red beds are tightly folded (General Geological Department of the Democratic Republic of Vietnam, 1973), which implies significant crustal shortening in the Mesozoic and Cenozoic.

In northeastern Thailand, ultramafic (serpentines, peridotites and pyroxenites with talc, magnetite and chromite pods) and mafic rocks (metagabbros, amphibolites, and metavolcanics of basaltic–andesite composition) are exposed for ~ 150 km along the NE–SW-striking Nan–Uttaradit suture zone (Barr and Mac Donald, 1987) (Figs. 1 and 2). Northeast of Uttaradit, these rocks, together with micaschists and black calcschists, form meters thick tectonic slivers separated by nearly vertical, NE–SW-striking, right-lateral shear zones. Blueschists, of epidote-blueschist facies, crop out a few kilometers to the west (Barr and Mac Donald, 1987). Along the southeast side of the zone, ultramafic rocks and sericite schists with large lenses of deformed andesites and polymict conglomerates, appear to have been thrust southeastwards onto a zone of steep, folded flyschs that locally display the typical Bouma sequence. The Nan–Uttaradit suture zone may continue into Laos to Luang Prabang and farther towards the NE, roughly parallel to the Dien Bien Phu strike-slip fault (Figs. 1 and 2) (e.g. Sengör and Hsü, 1984; Tapponnier et al., 1986, Hutchinson, 1989).

Whatever its northwards extension, that suture is usually interpreted to mark an important separation between the Kontum–Khorat and the Shan–Thai blocks. To the northwest, folded sedimentary and volcanic series include fossiliferous Lower Permian to Carnian–Norian shales, with sandstone and limestone intercalation, unconformably covered by Upper Triassic–Lower Jurassic acidic to intermediate volcanic effusives, tuffs, and volcanioclastics (Hess and Koch, 1975; Baum and Hahn, 1977; Hahn and Siebenhüner, 1982). These series have suffered at least two strong phases of shortening. One took place before the Upper Triassic, and the other, responsible
for relatively tight folds with axes trending N–S to NE–SW (Baum and Hahn, 1977; Department of Mineral Resources, 1982; Hahn and Siebenhüner, 1982) after the Jurassic–Cretaceous. The regional geology, and the fact that the Permo-Triassic Thai–Lincang granodiorite and the Nan–Uttaradit zone parallel each other (Fig. 1), has led most authors to interpret the southeastern margin of the Shan–Thai block as the site of northwest directed subduction in the upper Permian (in the present geographic coordinates) (e.g. Hutchinson, 1989; Barr et al., 1990, Sengör, 1984), followed by collision with the passive margin of the Kontum–Khorat block in the Middle Triassic (e.g. Bunopas and Vella, 1983; Sengör and Hsü, 1984). Helmcke (1985), however, among others, inferred that the collision occurred earlier. The fact that similar uppermost Triassic sediments and volcanic rocks lie unconformably on more deformed Paleozoic schists on either side of the Nan–Uttaradit zone (Baum and Hahn, 1977; Department of Mineral Resources, 1982; Hahn and Siebenhüner, 1982) supports the former view.

Small outcrops of reportedly ultramafic rocks are found at three places west of the Lincang batholith between Changning and Menglian in Yunnan (e.g. Bureau of Geology and Mineral Resources of Yunnan, 1983; Zhang Qi et al., 1985) (Figs. 2 and 3). These rocks, which are associated with Early Devonian–Middle Permian cherts and limestones, have been interpreted to outline a W-verging suture of Paleotethys, between the Bao–Shan and Simao blocks (Wu Haoruo et al., 1994). According to the latter authors, this suture continues into northwestern Thailand near Chiang Mai and Chiang Rai where basaltic rocks and, locally, ultramafics occur within the Thai–Lincang granitic belt (Department of Mineral Resources, 1982; Barr et al., 1990) (Fig. 2). On a petrological and geochemical basis, however, Barr et al. (1990) inferred that the Chiang Mai/Chiang Rai mafic and ultramafic rocks should be interpreted as the remnants of a back-arc basin located west of the ‘Nan–Uttaradit subduction’.

Toward the west, the Shan–Thai block is limited by Cenozoic fault zones: the Indoburman–Myitkyina suture, the Shan scarp, and the active Sagaing fault (e.g. Le Dain et al., 1984; Armijo et al., 1989) (Fig. 1). Southeast of the Nan–Uttaradit suture zone and Dien Bien Phu fault, the Khorat plateau undoubtedly remained more stable, particularly since the Upper Triassic. In the Upper Carboniferous and Permian, normal faulting affected the basement of the Khorat plateau (Mouret, 1994). Along the northeastern edge of the plateau, SW-verging thrusts, clearly imaged on seismic reflection profiles (Mouret, 1994), affect Upper Carboniferous–Lower/Middle Triassic sediments and are unconformably capped by Upper Triassic rocks (Indosian orogeny). During most of the Mesozoic and Paleocene, the Khorat basin formed a broad depocenter, comparable to the Sichuan basin, in which a thick sequence (up to 4500 m; Hutchinson, 1989) of mainly continental sandstones accumulated. That sequence was subsequently folded, probably in the Tertiary, into brachy anticlines and synclines oriented either roughly parallel to the NE–SW Nan–Uttaradit zone or to the NW–SE belts of Annam. Thus, as much of Yunnan, the Khorat suffered two roughly orthogonal shortening episodes after the Cretaceous/Paleocene. The orientations of shortening, however, appear to have been somewhat different, and the amounts much less.

As north of the Red River fault zone, the present-day deformation of northern Indochina is consistent with roughly N–S compression, but with more conjugate strike-slip faulting and only minor normal faulting (Fig. 5). Near the border between Burma and Yunnan, for instance, several NE–SW-trending faults that appear to be active on LANDSAT images (e.g. the Mengla, Mengxing, Jinghong and Nanting faults; Fig. 5) (Le Dain et al., 1984) offset, by 6–15 km, left-laterally, the Mekong and Salween rivers. On 6 November 1988, two earthquakes with $M_s = 7.6$ and 7.2 occurred near the northwestern tip of the Lincang fault (Fig. 5). The focal mechanisms of those shocks are compatible with right-lateral slip on the latter fault within a N–S compressional regime (Chen and Wu, 1989). In Vietnam, the NNE–SSW-trending Dien Bien Phu fault shows clear geomorphic evidence of active left-lateral slip, on satellite images, topographic maps, and in the field (Tapponnier et al., 1986).

The Red River Fault itself has long been known to be active, with a combination of right-lateral and normal faulting (e.g. Tapponnier and Molnar, 1977; Allen et al., 1984; Guo Shunmin et al., 1986; Tapponnier et al., 1990). As the longest and morphologically clearest fault in the area between Tibet and the
South China sea, it is also presumably the fastest slipping, with minimum slip rate estimates ranging between $\approx 3$ and 7 mm/yr (Allen et al., 1984; Guo Shunmin et al., 1986). The fault zone is rather linear and simple only between Midu and Lao Cai (Fig. 5), with at most two strands (Range Front and Mid valley Faults), a few kilometres apart (Fig. 9) (Allen et al., 1984). South of Lao Cai, in Vietnam, the fault splays into parallel strands along and north of the Day Nui Con Voi, and within the Tonghzing Alps. Motion appears to be oblique (normal–right-lateral) on several of those strands. North of Midu, the situation is even more complex. While oblique left-stepping, en-échelon normal faulting continues northwards, along either flank of the Diancang Shan range, followed by right-lateral faulting along the Qiaohou fault, bookshelf faulting on $\approx$ N–S-striking faults north of that range probably transfers right-lateral motion to the Zhongdian fault. The kinematics of faulting along the Red River and other regional active faults are nonetheless consistent with the regional stress orientations that characterize the current deformation of Yunnan (Fig. 5b) (Tapponnier and Molnar, 1977; Allen et al., 1984; Le Dain et al., 1984; Guo Shunmin et al., 1986; Duan Xinhua, 1981; Fan Chengjing, 1986). The Ailao Shan has thus long been interpreted to represent a suture between the Yangzi platform and Indochina. Various suturing ages have been proposed, ranging from Sinian (Precambrian; Fan Chengjing, 1986) to Paleozoic (Helmcke, 1985) or Indosinian (Middle–Upper Triassic; e.g. Li Chun-Yü et al., 1979; Bally et al., 1980). The steep limit between the gneisses and the schists, called the “Ailao Shan fault” (Figs. 9, 10, 11 and Fig. 12), has been interpreted to be a thrust, the gneiss belt thus forming an upthrust slice of the Precambrian basement of the Yangzi platform (e.g. Duan Xinhua, 1981; Cheng Yuankung, 1987). Hence, high-grade metamorphism in the Ailao Shan has generally been considered to be Precambrian. In this context, the existence of some Tertiary $K$/$Ar$ ages was only taken to imply retrograde Cenozoic metamorphism (Bally et al., 1980; Fan Chengjing, 1986; Wang and Chu, 1988). Tapponnier and coworkers have proposed a radically different interpretation (Tapponnier et al., 1986, 1990; Schärer et al., 1990, 1994; Zhong Dalai et al., 1990; Leloup, 1991; Le Dain et al., 1993; Leloup and Kienast, 1993), in which the Ailao Shan gneisses result from large-scale Cenozoic strike-slip shear, the schists and ultramafics having been smeared along the range by such motion. We present here most of the evidence that led us to this interpretation.

3. Tertiary deformation within the Ailao Shan–Red River metamorphic belt

In Yunnan, the Ailao Shan–Red River metamorphic belt is composed of two main massifs, the Ailao Shan and Diancang Shan (Figs. 2 and 3) (e.g. Academy of Geological Sciences of China, 1975). We describe below the structure, deformation, $P$–$T$ conditions, and age of metamorphism in these two massifs, targets of our most detailed field work.

3.1. The Ailao Shan metamorphic massif

The Ailao Shan metamorphic massif is less than 20 km wide and more than 300 km long. It juxtaposes two belts of rocks with contrasting metamorphic facies. Along the northeast side of the massif, high-grade gneisses make a continuous range that culminates at 3137 m (Figs. 3, 9 and 10). Lower-grade schists flank the gneisses along the southwest side of that range (Figs. 3, 9 and 10). Between Mojiang and the northern tip of the Ailao Shan, the schists contain small, dismembered bodies of ultramafic rocks elongated parallel to the range (Figs. 3 and 9). These rocks probably represent remnants of obducted oceanic crust and mantle. $HP/LT$ metamorphism is reported in the lower-grade schists (Huang, 1978; Bally et al., 1980; Duan Xinhua, 1981; Fan Chengjing, 1986). The Ailao Shan has thus long been interpreted to represent a suture between the Yangzi platform and Indochina. Various suturing ages have been proposed, ranging from Sinian (Precambrian; Fan Chengjing, 1986) to Paleozoic (Helmcke, 1985) or Indosinian (Middle–Upper Triassic; e.g. Li Chun-Yü et al., 1979; Bally et al., 1980). The steep limit between the gneisses and the schists, called the “Ailao Shan fault” (Figs. 9, 10, 11 and Fig. 12), has been interpreted to be a thrust, the gneiss belt thus forming an upthrust slice of the Precambrian basement of the Yangzi platform (e.g. Duan Xinhua, 1981; Cheng Yuankung, 1987). Hence, high-grade metamorphism in the Ailao Shan has generally been considered to be Precambrian. In this context, the existence of some Tertiary $K$/$Ar$ ages was only taken to imply retrograde Cenozoic metamorphism (Bally et al., 1980; Fan Chengjing, 1986; Wang and Chu, 1988). Tapponnier and coworkers have proposed a radically different interpretation (Tapponnier et al., 1986, 1990; Schärer et al., 1990, 1994; Zhong Dalai et al., 1990; Leloup, 1991; Le Dain et al., 1993; Leloup and Kienast, 1993), in which the Ailao Shan gneisses result from large-scale Cenozoic strike-slip shear, the schists and ultramafics having been smeared along the range by such motion. We present here most of the evidence that led us to this interpretation.

3.1.1. The high-grade gneisses

We were able to study the structure of the Ailao Shan, and the characters of ductile strain along seven sections spanning a length of 300 km (Fig. 9). Three sections cut the entire massif, with nearly continuous outcrops (sections $Ab$, $C$ and $Db$, Fig. 9). Evidence from 14 sites, spread over a distance of 90 km along the northeast flank of the massif, complement the sections (zones $E$ and $G$, Fig. 9). The fact that the northernmost section, located near the northern tip of
Fig. 9. Structural map of Ailao Shan massif with location of field observations in Yunnan (1983), modified according to field observations and LANDSAT data. Diagrams (lower hemisphere) show direction density diagrams of stretch. Maximum density areas are black. Local trend of metamorphic belt is marked by a cross-section of Fig. 12. B and F correspond to partial sections starting from river-polished outcrops, with each studied outcrop shown by star. The boundary of Ailao Shan is indicated by elongated crosses, which indicate deformed granite b.
field observations: drawn from Bureau of Geology and Mineral Resources of LANDSAT and SPOT satellite images analysis. Box a refers to Fig. 11. Schmidt net is of stretching lineations (L) and of foliations poles (S) in mylonitic gneisses. eic belt is shown by straight lines on each diagram. Ab, C and Db refer to ons starting at northern boundary of Ailao Shan. E and G correspond to zones n by star. H and J correspond to partial cross-sections starting at southern d granite bodies. M.V.F. = Mid Valley Fault; R.F.F. = Range Front Fault.
Quaternary and Neogene Gneiss  Norite
Mesozoic Low grade schists  Ultramafics
Paleozoic Granite
the Ailao Shan, lies 300 km away from the southernmost one (section Ab) close to the Vietnam border, and that neither geological maps nor satellite images reveal significant change of attitude or fabric along the gneiss belt, implies that our data set characterizes fairly well the overall structure and deformation of the belt.

3.1.1.1. Structure of the gneissic belt. High-grade gneisses form the elongated topographic backbone, 8–10 km wide core of the Ailao Shan (Figs. 9 and 10). The rocks include thinly banded, biotite-sillimanite–garnet-bearing paragneisses, orthogneisses, augengneisses with large feldspar porphyroblasts, migmatites, deformed leucocratic veins, and intrusions of anatectic leucogranites and granodiorites. The laminated paragneisses, which form a steep, continuous band along the northeastern border of the belt, are particularly well exposed in small canyons incising the steep range-front (zones E and G, Fig. 9). They contain stretched amphibolite layers and leucocratic veins, that form spectacular boudin trails, and rare boudins of garnet-pyroxenites, apparently derived from leached calc-silicates (Leloup, 1991;
Leloup and Kienast, 1993). Large, up to several hundreds meters wide boudins of intensely deformed marbles, with complex shapes, are found in the paragneisses, close to the border of the massif, along sections B, C, D, F and H (Fig. 12). Augengneisses, orthogneisses (deformed granites), and more or less deformed anatectic rocks are more common in the inner part and along the southwestern border of the high-grade core (e.g. sections C and Db, Fig. 12). Migmatites, some with complex flow folding, are exposed at places in the middle of the core. Light-gray, fine-grained gneisses (probably deformed rhyolites, or hypovolcanic granites), often with rod structure, are found along the southwestern part of the core along sections B, C and D. In zones E and G, and along sections B, C, D, and F (Figs. 9 and 12), the high-grade rocks form spectacular, steeply NE-dipping slabs, roughly parallel to the trend of the range. Between Mosha and Yuanjiang, such slabs are clear on SPOT images and on topographic maps, as they control the drainage direction of many streams (Fig. 11).

To the southeast, along section A (Figs. 10 and 12), paragneisses containing large amphibolite boudins and lenses of leucocratic gneisses maintain a steeply dipping foliation only near the northern limit of the core (sites A1, A2 and A3, Fig. 12). Along this section, a \approx 500 m thick band of recrystallized marbles, containing decimetric siliceous pods, is exposed 5 km south of the Range Front fault. South of these marbles, towards Yuanyang, the steeply foliated gneisses give way to flat-lying, altered micaschists embedding layers of gneisses, migmatites and pegmatites (Figs. 10 and 12). South of Yuanyang, the rocks dip again more steeply towards the north. Garnet-bearing micaschists containing lenses of leucocratic rod gneisses are found at the southern limit of the metamorphic core. Farther south, stripes of sedimentary rocks (mainly Paleozoic and Triassic), at most only slightly metamorphic, are affected by steep faults with structural evidence of left-lateral slip. Several boudins of marble (\approx 50 m thick) are found within the northernmost of these fault zones (Fig. 10).

### 3.1.1.2. Macroscopic characters of deformation.

Most rocks in the Ailao Shan gneissic core display pervasive mylonitic textures (Higgins, 1971; Bell and Etheridge, 1973; Hobbs et al., 1976). The small-scale mylonitic banding, the small size of matrix-grains and the strong deformation of porphyroclasts, which are often elongated and exhibit long tails, are qualitative indicators of the intensity of ductile strain (Figs. 13, 14, 15 and 16). The banding, defined by the strong elongation of layers with varied petrology, is particularly spectacular on stream-polished outcrops in zones E and G (Tapponnier et al., 1990), and along sections C, D and F (e.g. Fig. 13a,b). In the paragneisses of the northeastern border of the massif for example, thin dark layers of biotite-micaschist or amphibolite, alternate with layers bearing quartz–feldspar–biotite–sillimanite ± garnet. Deformed leucocratic veins are also part of the macroscopic banding. The different layers are generally affected by pervasive boudinage (Fig. 14).

The foliation, which is marked by the preferred orientation of planar minerals (micas) and by flattened quartz or feldspar ribbons, is parallel to the macroscopic compositional banding (Fig. 13d). The foliation planes bear a prominent stretching and mineral lineation (Fig. 13c). The lineation is marked by elongate quartz and feldspar ribbons, long tails of feldspar porphyroclasts, pressure shadows on garnets, and by the alignment of metamorphic minerals (biotite, amphibole, sillimanite). As in most strongly deformed rocks, the foliation plane and the lineation appear to be nearly parallel to the \( XY \) plane and to the \( X \)-axis of the finite strain ellipsoid (\( X \geq Y \geq Z \)), respectively (Flinn, 1965; Nicolas and Poirier, 1976). Most of the rocks display L–S tectonite fabrics, in which both foliation and lineation are well expressed, a fact suggestive of nearly plane-strain. Local, constrictional strains in leucocratic gneisses of the southwestern border of the Ailao Shan is attested.

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**Fig. 11.** Structural map of central part of Ailao Shan (boxed area on Fig. 9). Drawn from field mapping (bold drawing) and SPOT satellite image analysis (dotted lines). Elevations from 1:100,000 Chinese topographic maps, are in meters. **R.F.F.** = active Range Front Fault; **M.V.F.** = active Mid Valley Fault. **C** and **Db** refer to sections of Fig. 12.
Fig. 12. Geological cross-sections of Ailao Shan high-grade mylonites (located on Figs. 9 and 11). Arrows with numbers indicate observation sites and samples (Figs. 13–18). On C and Db, topography is drawn from Chinese 1:100,000 topographic maps; dashed line is projection of crest lines on sides of road section mapped in field. On section A, an approximate topography has been drawn from the Operational Navigation Charts at a scale of 1:1,000,000 and from field observations. Diagrams show foliation (great circles) and stretching lineation trends (arrows) in mylonitic gneisses, along each section (Schmidt diagrams, lower hemisphere). Arrows point towards shear direction (movement of hanging wall). Dashed line indicates overall direction of corresponding section.
by a rod fabric with strong lineation but ill defined foliation. Folds with axes parallel to the lineation (a-folds; e.g. Malavieille, 1987) often affect the mylonitic banding, while sheath folds (e.g. Quinquis et al., 1978) parallel to the lineation are locally observed (e.g. site G1, Fig. 9). Both type of folds are characteristic of progressive deformation in highly strained rocks (e.g. Cobbold and Quinquis, 1980; Lacassin and Mattauer, 1985; Malavieille, 1987).

The foliation, which is parallel to the slabs of gneisses visible on satellite images and in the landscape (Figs. 11 and 13e), is either parallel to the trend of the metamorphic belt or slightly more N-S striking (Fig. 9). Average foliation strikes are N124°E along section A, N133°E south of Yuanjiang (section B, Fig. 9), N128°E along section C, N149°E along section D, N146°E between Mosha and Ejia (zones E, G and section F), and N175°E east of Jingdong (section H) (Fig. 9). At the northern tip of the Ailao Shan, the foliation geometry seems to be rather complex with more scattered foliation poles. The foliation is generally steep with dips between 60 and 90° towards the NE (65° on average) (Figs. 9, 12 and 13). Along sections C and D, locally flatter foliations seem to be related to the complex three-dimensional geometry of large-scale marble boudins (Fig. 12). Migmatites also display foliations with various strikes and dips. In general, however, the foliation strikes parallel to the belt (Tapponnier et al., 1990).

The stretching lineation is in general nearly horizontal in the foliation plane and thus strikes nearly parallel to the metamorphic belt (Figs. 9 and 12). On sections A, C and D, the average lineations strike N110°E, N127°E and N148°E, respectively (Figs. 9 and 12). In the streams of the northeast flank of the Ailao Shan, it strikes N131°E between Mosha and Chunyuan (zone E), N140°E along section F, and N148°E between Chunyuan and Ejia (zone G). East of Jingdong, along section H, the lineation strikes N169°E (Fig. 9). The pitch of the lineation in the foliation plane is generally between 0 and 20° southwards. In zones where the foliation dips more gently (large-scale boudins of sections C and D, and the southern part of section A), the lineation remains parallel to the belt (Fig. 12) (Tapponnier et al., 1990). Rod lineations in leucocratic gneisses are also nearly horizontal and parallel to lineations in the surrounding gneisses. Despite some scatter, lineations at the northern tip of the Ailao Shan strike NW–SE to N–S.

3.1.1.3. Shear criteria. In all outcrops studied, we found impressive shear-sense indicators of various types (e.g. Berthé et al., 1979b; Simpson and Schmid, 1983; Hanmer and Passchier, 1991). At the outcrop scale, river-polished surfaces provide nearly horizontal sections close to the XZ plane of the finite strain ellipsoid (i.e. perpendicular to foliation and parallel to lineation). Such sections display particularly clear shear criteria. Asymmetric boudinage structures affecting amphibolite or leucocratic layers (e.g. Hanmer, 1986; Hanmer and Passchier, 1991), together with asymmetric foliation boudinage (Platt and Vissers, 1980; Gaudemer and Tapponnier, 1987; Lacassin, 1988), are widespread in the gneisses (Figs. 13a,b and 14). Left-lateral shear bands, typical of C–S or C'–S mylonitic fabrics (e.g. Berthé et al., 1979a, b), are also abundant, particularly in the orthogneisses of the southwest border of the core (Fig. 15) (Tapponnier et al., 1986, 1990). Feldspar porphyroclasts are usually strongly deformed, with elongate, asymmetric tails (Fig. 16c) that often form well-developed rolling structures (Fig. 16) (e.g. Lagarde, 1978; Simpson and Schmid, 1983; Passchier and Simpson, 1986; Van Den Driessche and Brun, 1987; Hanmer and Passchier, 1991). The wings or tails may be very long (typically 8 cm long for rotated porphyroclasts less than 1 cm wide), a fact that suggests very large shear strains (Van Den Driessche and Brun, 1987). It is often possible to observe rolling structures with clear stair-step geometry of the wings (Passchier and Simpson, 1986; Hanmer, 1990) mixed on the same outcrop with others with opposite wings lying in a single plane parallel to the shear plane ("in plane"; Hanmer, 1990; Hanmer and Passchier, 1991). Some isolated boudins also form rolling structures. In the augengneisses, feldspar porphyroblasts that grew during deformation often display snowball structures, reminiscent of those of garnets, with small rotated tails (Fig. 16b,e).

On each studied outcrop, a left-lateral shear sense, usually consistent with several of the above indicators, was clear. The different types of indicators were always mutually consistent. That the shear senses are consistent within each site and from one site to
Fig. 13. Deformation in the Ailao Shan shear zone. (a) Typical river polished outcrop displaying banded mylonites and migmatitic augen-gneisses with several generations of stretched leucocratic veins. Nearly vertical foliation strikes $\text{N}120^\circ\text{E}$; view towards N (site $\text{C}1$ on Figs. 11 and 12). (b) Laminated paragneiss mylonites at site $\text{G}1$ (Fig. 9); foliation dips steeply to NE and strikes $\approx \text{N}145^\circ\text{E}$, view towards NNW. (c) Sub-horizontal stretching lineation on nearly vertical foliation in gneisses; view toward SW (site $\text{A}2$, Fig. 12). (d) Polished slab of banded mylonite (sample YU42 from site $\text{C}2$, Fig. 12); nearly horizontal section, view from above, see corresponding quartz $c$-axes fabric on Fig. 18d. (e) Vertical amphibolitic paragneiss at site $\text{A}1$ (Fig. 9); foliation strikes $\approx \text{N}110^\circ\text{E}$; view towards SE.
another implies remarkably uniform kinematics at the regional scale. We were, in fact, unable to find a single outcrop with criteria opposite to the bulk left-lateral sense. Together with other inferences at a larger scale, discussed later, such remarkable consistency suggests that deformation was probably close to simple shear.

3.1.1.4. Boudinage and bounds on finite strain. Boudinage of centimetric to decametric scale is widespread, as a result of stretching of the layers that form the mylonitic banding. In the paragneisses, boudins of amphibolite often form trails tens of meters long. Several generations of leucocratic veins emplaced during the deformation (Leloup and Kien-
Fig. 14. Meter-scale boudinage in the Ailao Shan shear zone. (a) Stretched leucocratic vein at site C1 (Figs. 9 and 13a), view towards NW. Pencil gives scale. (b) Laminated paragneiss mylonites showing left-lateral shears. Stretched micaceous layers (insert) show disconnected boudins linked by elongated tails (site G1, Figs. 9 and 13b). (c,d) Stretched amphibolite layers, left-lateral shears separate clockwise rotated boudins; site G1 (Figs. 9 and 13b) and site D1 (Fig. 12), respectively. (e) Sketch of trails of amphibolite boudins at site G1 (Figs. 9 and 13b). Pervasive left-lateral shears and consistent asymmetry of boudins imply bulk left-lateral shearing. Initially adjacent boudins are often disconnected, with gaps larger than boudin size (see for example a and b); b–c are sub-horizontal outcrop surfaces viewed from above.
Fig. 16. Asymmetric deformation of feldspar porphyroclasts or porphyroblasts implying left-lateral shearing in Ailao Shan zone. All are nearly horizontal sections, viewed from above. (a,b,d,e) Examples of rolling structures at site C1 (Figs. 12 and 13a). Boxes show location of inserts. (c) Asymmetric tails on feldspar porphyroclast at site G1 (Figs. 9 and 13b). Coin gives scale. (f) Laminated paragneiss with asymmetric feldspar porphyroclast at site A2 (Fig. 12). Coin gives scale. (a) and (d) show well-developed rolling structures, which suggest large values of shear strain. On (a), note also the sigmoidal shape of the large feldspar porphyroclast in the center of photograph. (b) and (e) show synkinematic feldspar porphyroblasts in migmatitic augen-gneisses, forming incipient rolling structures.

Fig. 15. Left-lateral C–S and C'–S mylonitic fabrics in the Ailao–Shan shear zone. (a) Orthogneisses (site C2, Fig. 12), view towards NW. Lens gives scale. (b) Paragneisses (site D2, Fig. 12), horizontal section, view from above. Lens gives scale. (c) Migmatitic augen-gneiss with leucocratic bands (site G1, Figs. 9 and 13b), horizontal section, view from above. (d) Polished slab of sample YU43 (Site C2, Fig. 12).
ast, 1993) are also stretched. Boudins are generally asymmetric and often connected by left-lateral shears striking more easterly than the foliation plane (Fig. 14).

A detailed study of such amphibolitic and leucocratic boudins trails (Fig. 14) shows that initially adjacent boudins are usually well separated, with gaps often larger than the boudins themselves, suggesting very large strain (e.g. Gaudemer and Tappeynier, 1987). The restoration of such stretched layers using surface-balancing (Lacassin et al., 1993b) provides minimum layer-parallel extension $e$ (Ramsay, 1967):

$$e = \frac{(l_1 - l_0)}{l_0}$$

where $l_1$ is the length of the stretched layer, and $l_0$ is the restored layer length. The values thus obtained range between 590 and 870% for the amphibolite layers, and between 270 and 660% for the leucocratic veins (Lacassin et al., 1993b). Such values are only lower bounds for the total finite extension, since the layers first formed during the early stages of progressive deformation (transposition of the amphibolite bands parallel to the mylonitic foliation) or were emplaced late in the deformation regime (syn-tectonic leucocratic veins).

There are several instances of boudinage at a much larger scale (100 m to 1 km). Along section A, asymmetric foliation boudinage may be inferred to occur at a scale of at least 100 m, from detailed mapping of the foliation (Leloup, 1991) (Fig. 17). The asymmetry of such large-scale boudinage is consistent with left-lateral shearing. Kilometer-scale boudins of marbles are scattered along the northeastem flank of the gneiss core (Fig. 12). Careful study of existing regional geological maps (e.g. Bureau of Geology and Mineral Resources of Yunnan, 1983), LANDSAT and SPOT images also reveals that the southwestern border of the high-grade gneiss core is offset by left-lateral shear zones several tens of kilometers long (Leloup, 1991) (Figs. 9 and 11).

3.1.1.5. Microstructures and quartz c-axes preferred orientation. The study of the Ailao Shan mylonites at the microscopic scale confirms the observations made at larger scales. In thin sections, the foliation and the lineation are marked by metamorphic minerals (biotite, muscovite, chlorite, sillimanite, amphiboles) and by quartz or feldspar ribbons. Both the ribbons and the gneiss matrix are made of small crystals usually elongated in the foliation plane. Some minerals (e.g. feldspars, garnets, sillimanites), often elongated, have been fractured and stretched parallel to the lineation during late stages of the deformation.

Various types of shear criteria are observed in thin sections (Leloup, 1991; Leloup and Kienast, 1993). These criteria include (1) pervasive C or C' shear planes, (2) sigmoidal mica crystals (Eisbacher, 1970; Lister and Snoke, 1984), (3) feldspar porphyroclasts with very elongate asymmetric tails made of small, recrystallized quartz or feldspar, (4) rolling structures, and (5) asymmetric pressure shadows (e.g. Malavieille et al., 1982) of chlorite–muscovite on garnet porphyroclasts.

The quartz forms very long polycrystalline ribbons generally made of recrystallized grains (30–200 μm long on average). The grains have an irregular shape, sutured boundaries, often with undulose extinction, subgrains and deformation bands, attesting to internal deformation. In some samples, the matrix contains small elongated quartz grains with aspect
ratios of up to ≈ 10. Plastic deformation combined with dynamic recrystallisation thus appears to have been the dominant deformation mechanism in quartz (e.g. White, 1976).

Quartz c-axis fabrics, measured using the U-stage microscope, are well defined, with generally oblique girdles punctuated by several maxima (Fig. 18). Depending on the maxima positions, several types of fabrics may be distinguished: (1) fabrics with only one maximum on the Y-axis (e.g. sample GR3), (2) an oblique girdle with maxima close to the Y-axis (e.g. sample R17), (3) an oblique girdle with maxima in the XZ plane (e.g. sample R47), and (4) fabrics with crossed girdles (e.g. sample AY90). The maxima close to the Y-axis, found in most fabrics, suggests activation of the prismatic (a) slip system in many crystals (Burg and Laurent, 1978; Bouchez and Péccher, 1981). Such activation has often been considered indicative of relatively high-temperature conditions \( T > 400^\circ\text{C} \) (Blacic, 1975; Nicolas and Poirier, 1976). The maxima in the XZ plane, and slightly oblique to Z are typical of the basal (a) slip system (e.g. Burg and Laurent, 1978; Bouchez and Péccher, 1981). The fact that, in some samples, the c-axis fabric only displays one maximum rather than girdles, may suggest one predominant slip system (e.g. prismatic (a) in GR3, Fig. 18). Fabrics with oblique girdles, as in most samples, display constant asymmetry in the XZ plane consistent with left lateral shear (e.g. Laurent and Etchecopar, 1976; Burg and Laurent, 1978; Lister and Hobbs, 1980; Bouchez and Péccher, 1981; Etchecopar and Vasseur, 1987). The rare fabrics with crossed-girdles always show one predominant girdle and an asymmetric disposition of the c-axes maxima (e.g. AY90, Fig. 18). They are thus also consistent with the left-lateral shear regime.

One particular sample (YU29, Fig. 18) yields an unusual c-axis fabric with a maximum close to X and two secondary maxima in the XZ plane. The quartz c-axes were measured in one ribbon made of slightly elongated crystals of relatively large size (250 µm) with rectilinear or lobate boundaries. Since, according to Gapais and Barbarin (1986), such microstructures suggest high mobility of the grain boundaries, hence high temperatures during deformation, the fabric obtained on this sample might reflect activation of the prismatic \( <a> \) glide system (Gapais and Barbarin, 1986; Mainprice et al., 1986), which may only occur at very high temperatures (Nicolas and Poirier, 1976).

3.1.2. The low-grade schists

South of the Ailao Shan fault, the low-grade schist belt runs parallel to the gneiss core from the northern tip of the range to at least Mojiang (Figs. 3 and 9). There, the belt reaches a maximum width of ≈ 10 km (Fig. 3). The low-grade schists are significantly less deformed than the high-grade gneisses. The schistosity strikes are more dispersed than is the foliation in the gneisses. They range between N110°E and N180°E, with an average direction of N130°E along section C, and between N140°E and N200°E, with an average direction of N160°E, along section D. The schistosity is thus parallel to, or more northerly trending than the gneissic foliation. It is usually steep and rarely bears a stretching lineation. Where present, the lineation strikes parallel to the stretching lineation in the high-grade gneisses.

The largest bodies of ultramafic rocks, up to 2 km wide and 8 km long, are exposed in the central part of the schist belt, northwest of Mojiang and near Shuang Gou (Figs. 3, 9 and 11). They are mostly composed of serpentinized harzburgites and gabbros with ophitic structure. In general they are separated from surrounding upper Mesozoic sandstones by steeply dipping, cataclastic faults. On geological maps these elongate ultramafic bodies form trails parallel to the Ailao Shan and are separated by left-lateral faults like those offsetting the southeastern border of the gneissic core (Figs. 3, 9 and 11). Southwest of section C, near Anding, a small serpentinite massif lies in thrust contact with Mesozoic red beds. Near section D, the southwestern border of the low-grade schists corresponds to a spectacular alignment of small serpentinite bodies along a fault striking ≈ N135°E (Fig. 11). The serpentinite is affected by a steeply dipping, N170°E-striking, schistosity. The schistosity is in turn cut by two sets of steep shear planes (N30°E-striking dextral planes and, more numerous N135°E-striking left-lateral planes), indicating N75°E shortening. The obliquity of this direction to the N135°E-striking fault, and the predominance of left-lateral planes implies bulk left-lateral shear, parallel to the Ailao Shan range. We therefore
interpret these ultramafic bodies as large-scale boudins resulting from post-Mesozoic left-lateral shear along the range (Figs. 9 and 11).

3.1.3. Summary of deformation in the Ailao Shan

Foliations, stretching lineations, and shear criteria in the strongly deformed Ailao Shan high-grade
gneisses, coherent from the microscopic to the regional scale, suggest that the deformation regime was close to simple shear. The intense strain recorded in the rocks implies that the foliation lies at a low angle to the bulk shear-plane, and that shear occurred in a direction close to the stretching lineation (e.g. Escher and Watterson, 1974; Mattauer, 1975). The mylonitic rocks of the Ailao Shan thus formed in a large-scale horizontal left-lateral shear regime, parallel to the range, under high-temperature conditions.

When approaching the Ailao Shan core from the south, the regional N-S axial plane cleavage of the folded red beds of the Simao basin, blends with the cleavage in the low-grade schists, which progressively bends to become NW-SE, parallel to the foliation direction in the high-grade gneisses (Figs. 3, 9 and 11). We interpret this change to result from progressive deformation during a single phase of E-W shortening, responsible for both the N-S-trending folds with N-S schistosity in the Simao and Chuxiong basins, and for the NW-SE-trending mylonites within the Ailao Shan left-lateral shear zone (Tapponnier et al., 1986, 1990; Leloup, 1991).

3.2. The Diancang Shan massif

The Diancang Shan ("mountain of changing sky cloud") is a more bulky mountain massif than the Ailao Shan. About 15 km wide and 80 km long, it culminates at 4122 m. Its jagged crest towers above the surrounding country, which stands at an average elevation of 2200 m (Fig. 19). The Diancang Shan lies along the projected northwestward trend of the Ailao Shan (Figs. 2 and 3). The two massifs are separated by the "Midu Gap", an almost 80 km long stretch without metamorphic rocks (Fig. 3) (Tapponnier et al., 1990). The Diancang Shan forms a rising horst mostly composed of high-grade rocks, bounded on either side by en-échelon active normal faults (Figs. 5 and 19) (e.g. Allen et al., 1984; Leloup et al., 1993). Folded Mesozoic rocks fill the Yangbi basin to the west (Figs. 2 and 3), while mostly Paleozoic rocks are exposed to the east. Neogene and Quaternary deposits lie unconformably on the older, folded rocks on the hanging walls of the active normal faults. The largest Neogene-Quaternary basin is that occupied by Er Hai lake along the eastern flank of the Diancang Shan (Fig. 19).

We have studied three sections across the metapelitic rocks of the Diancang massif in some detail. The first one is along the river gorge west of Xiaoguan, near the southern end of that massif (P2 and P3, Fig. 19). The second one reaches the 4092 m high summit west of Dali (section Q, Fig. 19). The last section (section R, Fig. 19) is across the northern tip, only 2.5 km wide, of the massif near Qiaohou.

3.2.1. High-temperature deformation

The Diancang Shan metamorphics comprise rock types very similar to those seen in the Ailao Shan, including paragneisses, augengneisses, skarns, leucocratic melt layers, micaschists, hornblende schists and marbles. The Dali white and gray marbles have made that city famous in all of China for at least 1200 years ("Dali Shi" is the Chinese word for marble). Along the east flank of the mountain (section Q, Fig. 20), marbles, mylonitic paragneisses, orthogneisses and amphibole-bearing micaschists form spectacular, vertical or steeply ENE- or WSW-dipping slabs striking parallel to the massif (Figs. 20 and 21). The gneisses contain stretched leucocratic veins (generally pegmatites) and rare boudins of skarns. The western part of the section exposes W-dipping paragneisses, migmatitic gneisses and micaschists. At the southern end of the massif (P, Fig. 19).
19), the gneisses dip more gently towards the west or south. The northernmost section displays only a narrow stripe of fresh paragneisses, micaschists and calcschists.

3.2.1.1. Macroscopic characters of deformation. As those of the Ailao Shan, most Diancang Shan rocks show evidence of intense ductile strain. Widespread mylonitic textures include small-size matrix grains, strongly deformed porphyroclasts, and small-scale banding. The banding is parallel to a prominent foliation that bears a spectacular, nearly horizontal lineation. The foliation, marked by the preferred orientation of planar minerals (biotite, ± muscovite) and by the flattening of quartz or feldspar ribbons, is parallel to the large-scale gneiss slabs visible in the landscape and to the petrographic boundaries (Figs. 20 and 21a,c). The lineation is marked by the elongation of quartz or feldspar ribbons, by the tails of feldspar porphyroclasts in the gneisses, and by the alignment of metamorphic minerals (biotite, amphibole, tourmaline) (Fig. 21b,c,d). Locally, marbles and gneisses are affected by folds with nearly horizontal axes and by sheath folds (Fig. 21e).

In places, the high temperature (HT) foliations and lineations are overprinted by late, retrograde (LT) microstructures (Leloup et al., 1993). In particular, in the eastern part of section Q, the HT foliation bears striations marked by chlorite (e.g. at site Q1, Fig. 20b). West of Xiaguan, at the mouth of the river gorge (sites P1 and P2, Fig. 19), chlorite and sericite mark penetrative east-dipping foliation and lineation in the zone of LT mylonites that forms the eastern boundary of the massif (sites P1 and P2) cut across the HT foliations.

Despite the complexity of the foliation geometry, the trend of the HT lineations remains nearly parallel to the trend of the Diancang Shan (Figs. 19 and 22a,b,c). The HT lineations strike N163°E on average in the eastern part of section Q (Figs. 19 and 22a), and N165°E in its western part (Figs. 19 and 22a). Along section R (Fig. 19), the lineations strike ≈ N143°E (Fig. 22b). In zone P, the HT lineations, which strike N120°E on average, are slightly oblique to the trend of the range (Figs. 19 and 22c). They are statistically distinct from the LT lineations that strike N97°E on average at site P1, and N91°E at site P2 (Figs. 19 and 22d) (Leloup et al., 1993). The LT lineations have average pitches of 70°S on the LT foliation planes (Fig. 22d).

At one site along section Q (site Q4), we have been able to estimate the minimum finite elongation with the boudinage-restoration technique used in the Ailao Shan (Lacassin et al., 1993b). The restoration of syntectonic leucocratic veins at this site yields a mean layer parallel extension value of 540%. The occurrence of sheath folds (Fig. 21e), and of rolling structures and mylonitic textures, all suggestive of large strain, are in keeping with such large finite extension along this section.

3.2.1.2. Macroscopic shear criteria, microstructures, and quartz c-axes preferred orientation related to HT deformation. As in the Ailao Shan, we found ubiquitous evidence for non-coaxial deformation in most rock types of the Diancang Shan. Left-lateral shear criteria at the outcrop scale include C–S or C′–S mylonitic fabrics, elongated asymmetric tails and rolling structures next to deformed porphyro-
Fig. 20. Geological section across the Diancang Shan massif (section Q located on Fig. 19). Arrows with numbers show location of studied sites or samples. (a) Whole section. (b) Detailed section of upper part of eastern flank of Diancang Shan [box on (a)]. (c) Detail showing geometry of foliation between sites Q1 and Q3. (d) Sketch of 3-D geometry of foliation and lineation. Quartz c-axis preferred orientation of samples YU6 and D30 are shown on Fig. 24b and c.
Fig. 21. HT deformation in Diancang Shan. (a) Sub-vertical foliation in gneisses, view towards SSE of crest south of section Q (Fig. 19). (b) Vertical foliation in marbles with clear stretching lineation (pitch = 10°N) (site Q2 on Fig. 20). (c) Landscape showing slabs of steeply dipping gneisses with sub-horizontal lineation (pitch = 12°N), view towards NW (site Q1 on Fig. 20b). (d) Steeply dipping gneisses with nearly horizontal lineation at Diancang Shan summit, view towards N (site Q5 on Fig. 20b). (e) Eye folds and mushroom-like folds in paragneisses at site Q4 (Fig. 20b); such structures are likely to be sections of sheath folds. Knife gives scale.
clasts, and asymmetric foliation boudinage in micaschists (Fig. 23). At the microscopic scale, similar shear criteria also include deformed mica crystals of sigmoidal shape, often limited by C or C' shear planes. The quartz c-axis preferred orientations, which resemble those obtained in the Ailao Shan, also imply relatively high-temperature conditions (maxima close to Y). Left-lateral shear senses are confirmed by the asymmetry of oblique girdles (Fig. 24).

At each site along section Q (Fig. 20), the different types of shear criteria are mutually consistent and indicative of left-lateral shear. They are particularly outstanding on the eastern flank of the range (Fig.

Fig. 22. Lineations and foliations in Diancang Shan. Schmidt diagrams, lower hemisphere projection. Sections and zones are located on Fig. 19.
23) and scarcer, but unambiguous, on the western flank (site Q6). It is important to note that the shear-senses remain left-lateral all along the section up to the Diancang Shan summit, despite the change from E- to W-dipping foliations.

In the gorge west of Xiaguan (site P3, Fig. 19), consistent shear criteria (asymmetric tails of deformed feldspar porphyroclasts, C or C' shear planes) in vertical sections parallel to the HT lineation indicate top-to-the-SE shear, hence oblique (normal left-lateral) movement. The average pitch of the shear directions in the foliation planes is 56°E.

3.2.1.3. Variations in foliation dips at 100 m scale.
Along section Q (Figs. 19 and 20), a ~ 200 m wide zone with more gently dipping foliations may be followed in the landscape (Fig. 20c). This zone lies between the generally steep gneiss or marble slabs that make up much of the eastern flank of Diancang Shan (Fig. 20b,c). HT lineations in the gently dipping gneisses (~N155°E) are parallel to the range and to lineations in adjacent steeper gneisses. Asymmetric tails and rolling structures on feldspars indicate top-to-the-NW shear. Such shear is compatible with that observed in the steeper gneisses, and hence with left-lateral movement, in spite of the change in foliation dip (Fig. 20d). The complex structure of the zone may result from three-dimensional, asymmetric foliation boudinage during progressive left-lateral/normal shear (Fig. 20d). In that zone, as in much of section Q, the lineations have small, but constant pitches towards the north (10–20° on average; Figs. 21b and 22a), which is consistent with a normal component of movement on NE-dipping planes in a left-lateral regime (Fig. 20d). Similar structures, with changes in foliation dip without changes in shear direction, also exist along section C in the Ailao Shan (Fig. 12). In either range, such structures appear to be associated with marbles.

3.2.2. Low-temperature chlorite-bearing slickensides and lineations
The chlorite-bearing slickensides overprinted on the HT foliations in the eastern part of section Q are nearly down-dip (pitches ranging between 70°S and 70°N) on steep planes. Asymmetric chlorite crystallizations indicate normal movement, down to the ENE, on such planes, posterior to HT left-lateral shear (Leloup et al., 1993). Farther southwards, west of Xiaguan (sites P1, P2, Fig. 19), the LT chlorite-bearing mylonites, with E-trending lineations (Fig. 22d), display asymmetric feldspar porphyroclasts and asymmetric mica fish. Such shear criteria are also indicative of LT ductile normal shear (with a small right-lateral component) down to the east in this mylonite zone (Leloup et al., 1993). Both the LT mylonites and the chlorite-bearing slickensides appear to result from movement on the still active normal fault that bounds the Diancang Shan to the east. Such ductile to brittle structures probably have been progressively uplifted in the footwall of that fault (Tapponnier et al., 1990). The direction of extension consistent with the chlorite slickensides or with the LT lineations is compatible with the Quaternary strain field deduced from the active fault pattern in the area (e.g. Allen et al., 1984; Tapponnier et al., 1990).

3.2.3. Summary and discussion of deformation characters in Diancang Shan
The gneisses of the Diancang Shan core display spectacular evidence of intense high-temperature strain. Consistent shear criteria are found at most sites and, given the intense strain in the rocks, shearing must have occurred on planes nearly parallel to the foliation, in a direction close to the lineation (Escher and Watterson, 1974; Mattauer, 1975). Thus, as in the Ailao Shan, the gneissic core of the Diancang Shan appears to have been the site of large-scale, left-lateral, ductile shear in a ~N160°E direction parallel to the range. All the structural evidence suggests that the major deformation episode in both massifs is the same and that they both belong to a single crustal shear zone.

The southern termination of the Diancang Shan core (site P3, Fig. 19), shows more complex HT deformation geometry and kinematics than the center of the range. The fact that more gently dipping foliations wrap around that termination implies 3-D strain (Fig. 19). Top-to-the-SE shear, parallel to N120°E-trending lineations with both left-lateral and normal components of movement, has occurred there. Such shear is compatible with, if oblique to, the left-lateral shear senses within the steep inner core of the range (Fig. 19). We have inferred that the shape of the HT foliation at the southern termination of the
Fig. 23. Examples of shear criteria on the eastern flank of Diancang Shan. All pictures are details of outcrops perpendicular to HT foliation and parallel to HT lineation, viewed from above. (a,b) Rolling structures at site Q4 (Fig. 20b), scale given by coin. (c) C–S structures in micaschists at site Q3 (Fig. 20b,c), coin gives scale. (d) C–S structures at Diancang Shan summit (site Q5 located on Fig. 20b), scale is given by pencil.
Diancang Shan massif results from late dissection of the Diancang Shan–Ailao Shan shear zone by a large left-lateral/normal shear plane dipping to the SE (Leloup et al., 1993) (Fig. 25b). Such oblique motion would have contributed to unroof and uplift the gneissic core of the Diancang Shan, while burying the northern extremity of the Ailao Shan under the Midu gap.

3.3. P–T conditions and age of deformation in the Ailao Shan and Diancang Shan

A striking character of the Ailao Shan–Red River metamorphic belt is the rather sharp horizontal metamorphic gradient found on either side of its gneiss core. Unmetamorphosed rocks crop out between 3 and 10 km away from the high-grade gneisses (e.g.)

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Fig. 24. Quartz c-axis preferred orientations in Diancang Shan (Schmidt diagrams, lower hemisphere). Measurements made on U-stage are presented in XZ plane. L is lineation and X-axis. Contours 1, 2, 3, and 5%. Small Schmidt diagrams on lower right side of each c-axis fabric show attitude of foliation (great circle), lineation (dot), and thin section plane (dashed great circle), relative to geographic axes. (a) Sample D13 (located on Fig. 19). (b) Sample YU6 (section Q, located on Fig. 20a), 527 c-axes. (c) Sample D30 (section Q, located on Fig. 20b).
Bureau of Geology and Mineral Resources of Yunnan, 1983; Bureau of Geology and Mineral Resources of Yunnan Province, 1990; Tapponnier et al., 1990). Such steep thermal gradients might result from two causes: (1) high-temperature metamorphism restricted to the shear zone in close relation with the shear event, and (2) a large amount of differential uplift between the gneisses and the surrounding sedimentary rocks. Petrological (Leloup et al., 1993; Leloup and Kienast, 1993) and geochronological (Scharer et al., 1990, 1994; Harrison et al., 1992a, 1995; Liu et al., 1992; Leloup et al., 1993) data suggest that both processes took place.

Geological and structural evidence for high-temperature deformation in the gneisses is compelling (Leloup and Kienast, 1993). The banded parag-
neisses along the eastern flank of the Ailao Shan (i.e. zones E and G, Fig. 9) are amphibolite-facies rocks with biotite, garnet and frequent sillimanite defining the stretching lineation (Bureau of Geology and Mineral Resources of Yunnan Province, 1990; Leloup and Kienast, 1993). Some migmatitic rocks and leucocratic melts emplaced during progressive shearing are strongly deformed (with layer parallel extension of up to 660%; Figs. 13a,b and 14a) (Lacassin et al., 1993b). The common occurrence of prismatic ⟨a⟩ glide in quartz suggests temperatures in excess of 400°C during deformation. Furthermore, activation of prismatic ⟨c⟩ glide in YU29 (Fig. 18) implies, at least locally, temperatures close to the granitic solidus. Strong recrystallization of feldspar porphyroblasts in gneisses (Fig. 16) also testifies to high-temperature deformation.

The active Range Front normal fault (RFF) sharply bounds the Ailao Shan high-grade gneisses to the northeast. North of the fault, Mesozoic sediments are only slightly metamorphosed. Such metamorphic contrast appears to be due, in part, to uplift of the Ailao Shan gneisses along the RFF (Tapponnier et al., 1990). Toward the SW, although the Ailao Shan fault follows the boundary between the gneisses and the low-grade schists, the metamorphic gradient is less steep. Electron microprobe analysis performed on the blue amphiboles found in reportedly HP-LT rocks in those schists, has revealed either winchite fringed by actinolite rims (Zhang Ru Yuan et al., 1990), or magnesio-riebeckite (Leloup and Kienast, 1993), but no glaucophane. P–T conditions in the schists thus appears to have been only around 4 kb and 170°C (Zhang Ru Yuan et al., 1990).

A sharp metamorphic contrast exists on either side of the Diancang Shan. Since the mountain is bounded by two active normal faults, a significant part of that contrast is probably due to recent extensional uplift.

A detailed petrological study of pelite-derived gneisses in zone E (Fig. 9) confirms high-temperature coeval with left-lateral shearing (Leloup and Kienast, 1993). There were two successive, stable parageneses during deformation. The most prominent paragenesis (P1) is characterized by amphibolite facies assemblages (biotite, garnet, sillimanite, quartz, K-feldspar, plagioclase). The second one (P2), found only in cracks or pressure-shadows affecting the porphyroblasts of P1, shows retrograde minerals typical of greenschist facies (chlorite, muscovite, ± biotite). Left-lateral shear criteria are coeval with either paragenesis, implying a retrograde deformation path (cooling during shearing). Together with biotite–garnet and plagioclase–garnet thermobarometry, the compositional zoning of P1 garnets suggests a temperature increase during crystallization of the garnets from ~ 660°C and ~ 5 kb (P1a) to 710 ± 70°C and 4.5 ± 1.5 kb (P1b, close to the granitic solidus) (Fig. 26) (Leloup and Kienast, 1993). P2 P–T conditions are estimated to be near 500°C and under 3.8 kb (with large uncertainties) (Fig. 26). The metamorphic grade seems to be slightly lower in the Diancang Shan than in the Ailao Shan, but the scarcity of originally pelitic rocks precludes precise P–T estimates in much of the massif. So far, only one sample from the southern border of the massif (DC3, site P3, Fig. 19), a micaschist containing staurolite and garnets, has permitted P–T estimates. They imply nearly isothermal evolution from ~ 7 kb (550°C) to ~ 5 kb (570°C) (Fig. 26) (Leloup et al., 1993), with higher pressure and lower temperatures than in the Ailao Shan. The rocks exposed at the southern end of Diancang Shan were thus deformed deeper (between 25 and 18 km), albeit at lower temperatures than those along the northeastern flank of Ailao Shan (~ 18 km) (Fig. 26). In both cases however, the P–T estimates correspond to a relatively high geothermal gradient.

Rb–Sr ages obtained by Chinese geologists on metamorphic rocks of the Ailao and Diancang Shan mostly range between 840 and 1700 Ma, with a cluster between 840 and 900 Ma (Bureau of Geology and Mineral Resources of Yunnan Province, 1990; Fan Chengjing, 1986; Zhong Dalai, pers. commun., 1988). K/Ar ages on micas range between 11.5 and 230 Ma, with most of them comprised between 11.5 and 50 Ma (Wang and Chu, 1988). However, without a precise description of the sampling locations, samples, and of the analytical method used, it is difficult to assess which of these age groups, if any, relates to the deformation observed in the ranges. More recent geochronological results obtained on accurately located samples in the Diancang and Ailao Shan gneisses (Schärer et al., 1990, 1994; Harrison et al., 1992a, 1995; Liu et al., 1992; Leloup et al., 1993) yield a much clearer picture.
Fig. 26. $P-T$ evolution of Ailao Shan and Diancang Shan gneisses. Data from Leloup et al. (1993) and Leloup and Kienast (1993). $1a$, $1b$ and 2 correspond to successive parageneses observed in the paragneiss of the NE border of Ailao Shan (site G1 and zone E, Fig. 9). DC3 is a sample from zone P3 at the southern extremity of Diancang Shan (Fig. 19). As = aluminosilicate ($Ky$ = kyanite, $And$ = andalusite, $Sill$ = sillimanite); $Bi$ = biotite; $Cd$ = cordierite; $Ch$ = chlorite; $Ct$ = chloritoid; $Gt$ = garnet; $Kf$ = K-feldspar; $Mu$ = muscovite; $Qz$ = quartz; $St$ = staurolite.

U/Pb ages on zircons, monazites and xenotimes from leucogranitic veins, parallel to the foliation and affected by left-lateral shear are similar in both massifs (Fig. 27). In the central Ailao Shan, the ages obtained from two different veins (YS9, site E1 and YS11, section C) are $23 \pm 0.2$ Ma ($2\sigma$) (Schärer et al., 1990). New ages on zircon and titanite fractions from monzonites, monzosyenites and pegmatites
range between 26.3 ± 0.3 and 24.1 ± 0.2 Ma (Schärer et al., 1994). In the Diancang Shan, monazites, xenotime, and zircons from a leucocratic layer along section Q (YS35, Fig. 20b) yield ages between 24.4 ± 0.2 and 22.3 ± 0.3 Ma (2σ) respectively (Liu et al., 1992; Schärer et al., 1994). The longer time separating the monazites and xenotime closure ages probably reflect slower cooling than in the Ailao Shan (Liu et al., 1992), in keeping with P-T estimates. Such ages date the crystallization of the leucogranitic melts (Schärer et al., 1990, 1994; Liu et al., 1992). The U/Pb ages have been complemented by more than 40 new 40Ar/39Ar measurements on biotites, white micas, amphiboles and K-feldspar from the Ailao Shan and the Diancang Shan (sections A, C, D, E, F, G, P and Q) (Harrison et al., 1992a, 1995; Leloup et al., 1993). Thermal histories have been retrieved from K-feldspar age spectra using an extension of the single diffusion domain/activation energy closure theory (Dodson, 1973) that applies to minerals with a discrete distribution of domain sizes or activation energies (Lovera et al., 1989, 1991, 1993; Harrison et al., 1991; Fitz Gerald and Harrison, 1993). None of these studies yields ages older than 27 Ma. The 40Ar/39Ar measurements from samples collected close to the U/Pb dated samples (sections C, Q and zone E) systematically give ages younger than the U/Pb ones (Fig. 27) (Harrison et al., 1992a; Leloup et al., 1993).

The cooling histories of zone E, section C (Ailao Shan) and zone P and section Q (southern part of Diancang Shan) have been retrieved from K-feldspar thermal histories (40Ar/39Ar data), and from ages and closing temperatures of monazite and xenotime (U/Pb ages), amphibole, white mica and biotite (40Ar/39Ar ages) (Fig. 27). For each zone, results from different laboratories and with different techniques, all outline simple and coherent cooling histories. The ages corresponding to the higher temperatures (U/Pb ages) are significantly older than the end of the left-lateral shear episode. In zone E, the age obtained on the FA2 amphibole (Harrison et al., 1992a), whose closing temperature is comparable to that estimated for paragenesis P2 in the same zone (Leloup et al., 1993), suggests that left-lateral shear was still going on around 22 Ma (Fig. 27b). In zone E and section C (Ailao Shan), temperatures dropped below those compatible with quartz plasticity (≈ 300°C) around 20 Ma. This suggests that ductile shear came to an end at that time in the sampled rocks, although brittle left-lateral slip could have continued. In the Diancang Shan, temperatures of ≈ 300°C were reached around 17 Ma but maintained until about 4.7 Ma (Leloup et al., 1993) (Fig. 27c). More recent evidence suggest that cooling was diachronous along the belt. In fact, temperature dropped below ≈ 300°C at ≈ 22 Ma on section A, at ≈ 20 Ma on section C, at ≈ 19 Ma on section D (Fig. 27a), and at ≈ 18 Ma for sections E, F and G (Harrison et al., 1992a, 1995; Leloup et al., 1994) (Fig. 27a,b).

Hence, with the new evidence at hand, the overall picture that emerges is as follows. The left-lateral shear episode that formed the Ailao and Diancang gneiss cores is neither of Precambrian, nor of Mesozoic, but of Oligo-Miocene age. It must have started after the lower Tertiary (50–60 Ma), since Upper Cretaceous–Paleocene red beds are affected by kinematically consistent strain, and in all likelihood before 26.5 Ma. Ductile shear was still in progress at 22 Ma. Fast uplift subsequently caused the temperature to drop and ductile shear to stop around 20 Ma in the gneisses now exposed at the surface, even though, at the same structural level, brittle left-lateral faulting probably continued along zones now eroded, buried or juxtaposed with the active Red River Fault traces. The change to the present day stress field (NNW–SSE shortening) and the onset of right-lateral slip on the active Red River Fault took place later (Allen et al., 1984; Tapponnier et al., 1990), possibly around 5 Ma (Leloup et al., 1993).
cooling history (Fig. 27c) suggests two successive uplift episodes (Leloup et al., 1993). Rapid cooling starting at 4.7 Ma may correspond to the activation of the normal fault bounding the range to the east. Such faulting could account for \( \approx 10 \) km of exhumation and the formation of 4–5 km of structural relief (Tapponnier et al., 1990; Leloup et al., 1993).

Cooling between 24 and 17–19 Ma (Fig. 27) could correspond to movement on, and denudation of the Diancang Shan by, the large oblique shear plane that bounds the massif to the south (Leloup et al., 1993). This movement produced large-scale boudinage of the shear zone during late stages of left-lateral shear. It might also account for about 10 km of uplift.

In the Ailao Shan, given the maximum pressure estimate (paragenesis Plb, \( 5 \pm 1.5 \) kb), the gneisses have been uplifted less, about 18 km since crystallization of that paragenesis. The cooling history found in zone E implies that temperatures lower than those corresponding to P1 (\( \approx 700^\circ \text{C} \)) had been reached at \( \approx 23.5 \) Ma (Fig. 27b). If uplift had been uniform since 23.5 Ma, it would have occurred at a minimum average rate of the order of \( 0.8 \) mm/yr. The small pressure drop between Plb and P2 suggests that uplift started at the time of P1, during left-lateral deformation (Leloup and Kienast, 1993). Although such a pressure drop is within thermobarometers uncertainties, the fact that rapid cooling (\( \approx 110^\circ \text{C} / \text{Ma} \), Fig. 27b) took place at this time supports this inference. Along section C and in zone E, the cooling histories retrieved from the \(^{39}\text{Ar} / ^{40}\text{Ar} \) data (amphiboles and K-feldspars) show only one cooling phase (Fig. 27a,b) (Harrison et al., 1992a, 1995) apparently roughly coeval with the older cooling phase in the Diancang Shan (Fig. 27). In contrast, however, temperatures in the Ailao Shan dropped below \( 200^\circ \text{C} \) at \( \approx 19 \) Ma, \( \approx 100^\circ \text{C} \) less than in the Diancang Shan at the same time, and the thermochronological techniques used do not give information on the thermal history below \( 200^\circ \text{C} \).

Since the ductile strain and left-lateral offsets appear to be large, small deviations in the shear direction from the horizontal can induce significant vertical movement along parts of the shear zone (Tapponnier et al., 1990). This led Harrison et al. (1992a) to interpret rapid cooling around 20 Ma as a result of unroofing, due to a component of normal throw and oblique extension during the overall left-lateral shear regime. Reconstruction of the opening of the south China Sea, which implies that left-lateral movement on the ASRR zone results in transtension along the southeastern part of that zone, southeast of the Diancang Shan (Briais et al., 1993), is consistent with this interpretation.

Although the cooling history of the Ailao Shan after 19 Ma has not been deciphered yet, significant uplift of its gneiss core must have occurred on the active Range Front Fault, which bounds it to the east, even though the present structural relief (2–2.5 km), a lower bound of such uplift, appears to be less than that in the Diancang Shan.

### 3.5. Separation of the Ailao Shan and Diancang Shan: origin of the Midu gap

Although the same left-lateral shear episode is clear in both the Ailao Shan and Diancang Shan, which are elongated and aligned parallel to the shear direction, the two ranges are separated by the \( \approx 80 \) km long Midu gap, in which no metamorphic rock is exposed (Fig. 3) (Bureau of Geology and Mineral Resources of Yunnan, 1983; Tapponnier et al., 1990). The two ranges are connected across that gap by the active fault traces that form the northwest extension of the Red River fault zone (Fig. 5) (Allen et al., 1984). Between the southern tip of the Midu basin and section M east of Nanjie (Figs. 3 and 28a), those traces are lined with 1–2 km wide zones of dark,
clay-rich gouges (Leloup, 1991). Southwest of the gouges, Mesozoic red beds are strongly folded. Ten kilometers south of the fault zone, these steep sandstone beds exhibit no cleavage. Closer to it, they become more deformed and display spectacular, steep, fracture cleavage striking N140°E. Between those schistosed sandstones and the gouges, a ≈ 1.5 km wide stripe of calcschists affected by strong flow cleavage is exposed (Fig. 28a). At site M1 (Fig. 28a), the flow cleavage strikes N−S (± 20°) and is cut by two sets of conjugate shear planes consistent with N100 ± 10° directed shortening (Fig. 28b,c,d). The more prominent shear planes are left-lateral and strike N170°E on average (Fig. 28b,c). Minor, right-lateral shear planes strike N50°E on average (Fig. 28c). The progressive increase in strain intensity towards the NE, the size of the gouge zones, the predominance of left-lateral shears and the shortening direction are all consistent with large left-lateral movement, rather than with recent right-lateral slip (Allen et al., 1984) along the RRF zone. We conclude that the left-lateral Ailao Shan–Diancang Shan shear zone crosses the Midu gap, but at a higher structural level. North of section M, the brittle left-lateral shear zone, and the corresponding gouges appear to lay buried under the Quaternary fill at the edge of the Midu Basin (Leloup, 1991). This brittle zone is thus not strictly aligned with the Diancang Shan (Fig. 3). Such observations led Leloup et al. (1993) to propose the following structural history: (1) until ≈ 23 Ma, the Diancang Shan and Ailao Shan formed a continuous shear zone; (2) at least ≈ 20 km of left-lateral movement subsequently took place along the oblique shear plane at the southern tip of Diancang Shan between ≈ 23 and ≈ 17 Ma, accounting for the left step between that tip and the Midu brittle shear zone (Fig. 25b) and for ≈ 10 km of denudation and uplift of the Diancang Shan; (3) tectonic activity along the ASRR was minor, at least in this area, between ≈ 17 and 5 Ma; (4) starting at 4.7 Ma, slip on the active faults now bounding the Diancang Shan, leading to further rapid uplift of this massif and right-lateral offset of parts of the left-lateral shear zone, followed activation of the RRF in a right-lateral sense. This scenario accounts both for the two distinct uplift phases documented in the Diancang Shan and for the different structural levels now exposed along this part of the ASRR–RRF zone.

3.6. Could the Ailao Shan and Diancang Shan gneiss cores result from processes other than left-lateral shear?

When taken at face value and in its simplest acceptance, the evidence presented in this paper is compatible with the interpretation first proposed by Tapponnier and coworkers (Tapponnier et al., 1986, 1990), namely, that the Ailao Shan–Diancang Shan gneisses mark the trace of a large, left-lateral Tertiary shear zone between Indochina and South China. Is this interpretation uniquely constrained by the data, however? Could other, more complex scenarios account equally well for the observations? In particular, could the present attitude of the gneisses be very different from when they formed? Might they not correspond to a deformed décollement between basement and shortened sedimentary cover or within the middle–lower crust?

In that case, initially flat-lying gneisses (Fig. 29a) would have been subsequently bent into a steeper, apparently strike-slip attitude. In the Diancang Shan, along section Q, the foliation planes do display an inverted fan geometry (Fig. 20), with a common intersection parallel to the mean lineation (Fig. 22a). This geometry could reflect folding of an initially flat foliation about an axis parallel to the lineation. But that mechanism would require structural symmetry, and inversion of the shear criteria from one limb of the fold to the other (Fig. 29b). This can be ruled out, since the observed shear senses are left-lateral irrespective of the dip of the foliation. The inverted foliation fan thus probably reflects upwards thinning of the shear zone.

Warping by drag of a flat foliation near the normal faults that bound the gneisses could also account for their present attitude (Fig. 29c). This, however, would also result in dip-dependent shear senses: left-lateral along NE-dipping faults (Ailao Shan, east flank of Diancang Shan), right-lateral near SW-dipping faults (western flank of Diancang Shan), which is not observed (Fig. 29c). Besides, in the Ailao Shan, there is no evidence that the gneisses plunge under the low-grade schists towards the southwest.

The Ailao Shan gneisses could be interpreted as initially flat mylonites upthrusted onto the schists along the Ailao Shan fault (Fig. 29d). There is,
Fig. 28. Deformation in Midu gap, between Ailao Shan and Diancang Shan. (a) Cross-section M (located on Fig. 3); site M1 refers to photograph and sketch on (b) and (c). (b) Left-lateral shears in calcschists at site M1, shear planes and foliation strike N140°E and N20°E, respectively. Horizontal plane, view from above. (c) Sketch of conjugate shear bands observed at site M1. Dominant left-lateral shears imply rotational deformation. (d) Shaded zones represent respective ranges of observed left- and right-lateral shears. N100°E shortening direction inferred from conjugate shears is compatible with left-lateral shear along N140°E-trending ASRR zone.

However, little strain evidence for large-scale overthrusting (dip slip lineations, top-to-the-SW shear criteria) in the Ailao Shan, and upthrusting cannot account for the structure of the Diancang Shan.

All alternative hypotheses discussed above would imply warping of a flat décollement about an axis parallel to a preexisting lineation for a length in excess of 500 km, which seems quite improbable.
Fig. 29. Alternative hypotheses to interpretation of Ailao Shan and Diancang Shan as left-lateral shear-zones. All hypotheses involve a regional décollement level (a), later folded (b), affected by normal faults (drag folding), (c) or by thrust (d).

Such a lineation would trend NW–SE, a direction markedly different from that of the E–W shortening compatible with folding of the cover rocks, or that obtained by unfolding the lineation on the Yulong Shan décollement (Lacassin et al., 1993a, 1995). The sense of transport, for at least 500 km, would be towards, rather than away from the India–Asia collision front, also an improbable circumstance. We conclude that the interpretation of the Ailao Shan–Red River zone as a major left-lateral shear zone is both the simplest and most plausible one.

3.7. Continuation of the Ailao Shan–Red River shear zone towards the N and S

3.7.1. The Xuelong Shan and Sanjiang fault zone

3.7.1.1. The "Madeng gap" and the Xuelong Shan.

A ~ 90 km long hiatus separates the northern tip of the Diancang Shan from the northernmost high-grade metamorphic massif identified so far along the RRF, the Xuelong Shan (Fig. 6). By analogy with the Midu gap, we refer to this zone as the "Madeng gap". This gap is crossed by the trace of the active Qiaohou fault, along which Mesozoic calc-schists exhibit strong schistosity (section S, Figs. 3 and 7a), as on section M in the Midu gap (Figs. 3 and 28).

North of section T', west of the Weixi basin and east of the N–S Mekong valley, the Xuelong Shan ("Mountain of the snow dragon") forms a NW–SE-directed range about 50 km long and 10 km wide, culminating at 3909 m (Fig. 30a). As in the Ailao and Diancang Shan, the steep NE topographic flank of the range (Fig. 30b) is marked by conspicuous triangular facets attesting to recent uplift along the active normal fault bounding the Weixi Neogene–Quaternary basin. Until recently, the rocks of the Xuelong Shan had been mapped as slightly metamorphosed Paleozoic sediments (Bureau of Geology and Mineral Resources of Yunnan, 1983), possibly because the area was remote and the range, covered with abundant vegetation, not accessible by road. More recent geologic maps (Bureau of Geology and Mineral Resources of Yunnan, 1984) show that this range is made of NW–SE-striking, metamorphic strips: epimetamorphic Triassic rocks (mostly schists and marbles) to the east, and high-grade gneisses (gneisses and amphibolite schists) to the west (Figs. 7b and 30a). East of the Weixi basin, Triassic volcanics and sedimentary rocks are sliced by NW–SE faults, parallel to the range (Fig. 30a). Toward the west, N–S folds affect Mesozoic–Tertiary red beds (Figs. 6 and 30a).

We were able to study the deformation of the metamorphic rocks along an incomplete section only ~ 4 km long, near the southern tip of the massif (section T', Fig. 30). Approaching the range from the north, the Triassic volcanics and limestones are tightly folded, steeply dipping, strongly brecciated and sliced by cataclasite zones. A 40 m wide gouge
zone separates the massif core from these deformed rocks. This gouge zone, which is aligned with the steep NE flank of the range, probably links the recent normal fault along it with the Qiaohou fault farther south (Fig. 30a,b). West of the gouge zone, one finds schists, strongly deformed rhyolites, fine-grained gneisses, micaschists with leucogranite and amphibolite boudins, and migmatitic gneisses (Fig. 30b). All the metamorphic rocks display mylonitic texture and evidence of intense non-coaxial deformation. The foliation dips 37°E and strikes N161°E on average, nearly parallel to the N150°E strike of the Xuelong range (Fig. 30c). A strong stretching lineation is observed along the whole section (Fig. 31a), striking N170°E on average with a small pitch to the north (Fig. 30c). As in the Ailao and Diancang Shan, numerous shear-sense indicators are visible, including (1) C–S structures (Fig. 31b,c), (2) porphyroclasts with asymmetric tails (Fig. 31d), (3) asymmetric foliation boudinage, and (4) rolling structures. All these indicators are consistent with top (east side)-to-the-north or left-lateral shear.

Our preliminary study of the Xuelong Shan metamorphic rocks thus implies a structure and strain regime similar to those described along the Ailao and Diancang Shan. The temperature during deformation may also be comparable since deformed migmatitic paragneisses crop out at the southeast end of section T'. One notable difference is the attitude of the foliation, which is flatter than in most of the Ailao and Diancang Shan, with the exception of regions near their southern ends (e.g. section A, Fig. 12, and zone P3, Fig. 19). It is possible, however, that the foliation becomes steeper in the inner core of the central and northern Xuelong Shan, as suggested by the rather straight petrographic boundaries on either side (Fig. 30a) (Bureau of Geology and Mineral Resources of Yunnan, 1984). Like the other two ranges, the Xuelong range is bounded to the east by an active normal fault that probably accounts partly for the uplift of the gneisses. For these reasons, we conclude that, in all likelihood, the Xuelong Shan belongs to the Ailao Shan–Red River shear zone. Preliminary Eocene–Oligocene U/Pb ages (Schräer et al., in prep.) support this contention.

3.7.1.2. The Sanjiang fault zone. There appears to be no other comparable metamorphic massif northwest of the Xuelong Shan, at least in Yunnan, on extant large-scale geologic maps (e.g. Bureau of Geology and Mineral Resources of Yunnan, 1984; Bureau of Geology and Mineral Resources of Yunnan Province, 1990), although a N–S zone of “dynamo-metamorphism” reportedly exists along the upper Mekong valley, approximately between 27°40'N and 29°N (Wang and Chu, 1988, fig. 2). North of the Xuelong Shan, the Sanjiang (three rivers) area is densely sliced by N–S-striking fault zones (Fig. 6). One such prominent zone passes west of Deqen, at the western extremity of section U2 (Figs. 6 and 8a). There, vertical, kilometric slivers of various rocks are separated by gouge and cataclasite zones possibly associated with strike-slip movement. The sense of slip on that fault is not known, however, and both the Benzilan suture, and the Gaoligong Shan gneiss belt along the Salween valley (Zhong Dalai, pers. commun.), which strike N–S, show evidence of right-lateral shear. Besides, at the rim of the east Himalayan syntaxis, strike-slip deformation in the three rivers area has probably been overprinted by intense shortening. Thus, at this stage of our study, it seems equally possible that the RRF continues north of Xuelong Shan past Deqen, or that it has been offset right-laterally by the N–S-striking fault zones that follow the Mekong and Salween for hundreds of kilometers.

3.7.2. The Day Nui Con Voi

South and east of the Ailao Shan, the Day Nui Con Voi ("elephant back mountain range") stretches for about 300 km along the north bank of the Red River (Fig. 2). With a width of 10–15 km, it forms the southernmost massif of the ASRR metamorphic belt. In southern Yunnan, that range nearly merges with the Ailao Shan (Bureau of Geology and Mineral Resources of Yunnan, 1983), being separated from it by the active Red River valley fault (Fig. 2). Near the Yunnan–Vietnam border, this fault splays into two roughly parallel strands, the Song Chay and Song Hong faults, which bound the Day Nui Con Voi to the north and south, respectively (Figs. 5 and 32a). Currently, both fault-strands appear to slip mostly right-laterally, with variable components of normal slip.

As for the Ailao Shan, the Day Nui Con Voi metamorphic rocks (mostly gneisses) are mapped as
Fig. 30. Structure of Xuelong Shan. (a) Geological and structural map of Xuelong Shan massif. Drawn from Bureau of Geology and Mineral Resources of Yunnan (1984), modified from LANDSAT satellite image analysis and field work. Area of map corresponds to box c on Fig. 6. Section T' is shown on (b). (b) Geological section across southern tip of Xuelong Shan. Arrows with numbers show location of photographs of Fig. 31. (c) Schmidt diagrams (lower hemisphere) of foliation (line) and lineation (dots) along section Ob.
Proterozoic (General Geological Department of the Democratic Republic of Vietnam, 1973; Bureau of Geology and Mineral Resources of Yunnan, 1983). South of the Song Hong fault, the Ailao Shan terminates in the reportedly Paleogene Phang Si Pang granite massif (General Geological Department of the Democratic Republic of Vietnam, 1973). The only available geochronological study of that granite has provided K/Ar ages between 80 and 29 Ma (Izokh et al., 1964). Our preliminary mapping and sampling of the Day Nui Con Voi gneisses across one nearly complete section and at four sites outside that section shows that most rock types are mylonitic with a well-defined foliation, striking nearly parallel to the belt, as in the Ailao Shan (Fig. 32b). Along the Bao Yen section, the foliation is generally steep, dipping either towards the SW or NE, as in the Diancang Shan (Fig. 32a,b). As in the Ailao and Diancang Shan the stretching lineation is everywhere well defined and nearly horizontal (Fig. 32b). Shear criteria are left-lateral irrespective of the foliation dip (Fig. 32c,d). Preliminary U/Pb and \(^{39}\)Ar/\(^{40}\)Ar radiometric dating yields Eocene–Oligocene ages (Schärer and Leloup, work in progress).

Hence, we interpret the Day Nui Con Voi to be part of the ASRR shear zone, which brings this narrow belt of Tertiary, left-laterally sheared, metamorphic rocks to a total outcrop length of 900 km, almost to the shore of the South China Sea (Figs. 1 and 2).

4. Summary and discussion

4.1. The Ailao Shan–Red River shear zone, a reference model for HT lower-crustal strain in plate-scale strike-slip faults

The evidence for such movement at that time is both impressive and remarkably consistent, for nearly 1000 km along strike, throughout Yunnan and north Vietnam. Though misinterpreted or overlooked, surprisingly, by previous workers (e.g. Bally et al., 1980; Duan Xinhua, 1981; Allen et al., 1984), such evidence is magnificently exposed in the \(\approx 10\) km wide cores of four elongate ranges (Xuelong Shan, Diancang Shan, Ailao Shan, and Day Nui Con Voi) in which high-grade, amphibolite-facies gneisses stand exhumed.

Together with the aspect ratio \((1/100)\) of the metamorphic zone, all the measurements at hand rule out an origin by processes other than longlasting, localized strike-slip motion. The gneisses were deformed at depths of 15–25 km under peak temperature of \(710 \pm 70^\circ\)C and \(\approx 7\) Kbl, respectively. The steep, usually N-dipping, mylonitic foliations bear near horizontal stretching lineations everywhere parallel to the local trend of the zone. Left-lateral senses of ductile shear are ubiquitous and clear. Crystallization ages and cooling histories of various minerals in the gneisses demonstrate that such shear went on unabated for at least \(\approx 10\) Myr (between \(\approx 27\) and \(17\) Ma) (Schärer et al., 1990, 1994; Harrison et al., 1992a; Leloup et al., 1993). Since the location, chemical composition and age \((35 \pm 0.1\) Ma) of the Jianchuan monzonites imply that they derive from anatectic in the shear-zone continuation under the Madeng gap, the duration of shear was more likely longer than \(18\) Myr \((\approx 35–17\) Ma) (Schärer et al., 1994).

There is no geochronological evidence for magmatism and metamorphism during the Paleozoic and Mesozoic along the Ailao and Diancang Shan, as ought to be the case if the ASRR zone had been the site of subduction followed by suturing and collision during the Indosinian orogeny (e.g. Bally et al., 1980; Duan Xinhua, 1981). Regressions of discordant zircon ages only yield upper intercepts between 670 and 1600 Ma, implying a Proterozoic protolith for the gneisses (Schärer et al., 1994). Hence, contrary to that fashionable view (e.g. Dewey et al., 1989; Duan Xinhua, 1981; Wu Haoruo et al., 1994), and no matter how profound a geologic, tectonic and geomorphic discontinuity, that zone is not a Phanerozoic suture. In keeping with this conclusion, the dismembered ultramafics and low-grade schists ex-
posed along the southern Ailao Shan can only be remnants of a suture transverse to it, offset and smeared a long distance by transcurrent motion (Tapponnier et al., 1986, 1990). The final effect of strike-slip shear and "parasitic" uplift has in fact been to juxtapose two belts of metamorphic rocks with highly contrasting grades: one composed of recent, high-temperature rocks generated in the lower crust from an ancient protolith and locally uplifted, the other made of older low-temperature rocks transported at shallow depth in the upper crust. Such a simple mechanism probably accounts best for the formation of paired metamorphic belts in which lineation is parallel to trend.

To date, not only is the ASRR shear zone the only prominent zone of well-documented Tertiary metamorphism in Asia outside the Karakorum and Himalayas, but it appears to be the only exposed strike-slip shear zone of that size and age, with that degree of metamorphism and anatexy, worldwide. Among other zones with a similar strain style, that which follows the Alpine Fault of southern New Zealand bears perhaps the closest resemblance, but it seems less developed and is much less exhumed. The alpine schists, which are interpreted to derive from trench greywackes (Landis and Bishop, 1972), form a narrow (10–15 km wide) metamorphic stripe along the central segment of that fault, mostly between Mount Aspiring and Arthur’s pass. That distance, about 300 km, is shorter than the length of just the Ailao Shan in Yunnan. The schists appear to have reached only the lowermost amphibolite grade (garnet–oligoclase, $T < 550\,^\circ C$), and this mostly within $\approx 5$ km south of the fault. Although some mylonites with lineations parallel to fault-strike, and isolated, concordant pegmatite dykes are found adjacent to the fault ($\leq 1–2$ km) (Grindley, 1963; Reed, 1964), there is no description of the steep, many kilometers thick, horizontally lineated, mylonitic gneiss slabs, or of the widespread synkinematic intrusives and often pervasive anatectic melting that prevail in the cores of the Ailao and Diancang Shan. Originally, most of the strain and metamorphism of the alpine schists had been interpreted to have taken place during the Mesozoic (Rangitata orogeny, between 140 and 80 Ma), concurrent with $3/4$ of the total dextral displacement (480 km) that the offset between the Otago and Marlborough arc terranes implies on the fault (Wellman, 1955; Wellman and Cooper, 1971; Scholz et al., 1979). Only slight retrograde metamorphism, together with at most $\approx 12$ km of uplift, was reported to have accompanied the remaining horizontal displacement (120 km) in the late Cenozoic (Kaikoura orogeny, 5 Ma–present). Shear heating during the latter event was interpreted to have been enough to reset K–Ar and Rb–Sr ages near the fault (e.g. Sheppard et al., 1975; Scholz et al., 1979). More strain is now thought to have occurred since the mid-Tertiary, in better agreement with plate tectonic reconstructions (e.g. Weissel et al., 1977; Stock and Molnar, 1987). But where the Alpine schist belt is transcurrent, it is clear that such strain has not exposed hundreds of kilometers long, tens of kilometers wide massifs of mid-lower-crustal rocks that suffered prograde, high-temperature Tertiary strike-slip shear.

Data from the Ailao Shan–Red River zone thus best document what ought to be the typical thermo-mechanical regime and style of deep-crustal strain in large continental strike-slip faults, for which it provides a reference example. The recent age and high grade of the zone, in particular, has made it possible to begin dating thermal changes resulting from transcurrent deformation with unprecedented accuracy (Fig. 27).

The most constant structural character is the trend of the lineation, which is locally parallel to relative movement, everywhere and irrespective of the more variable dips of the foliation. Although such mylonitic foliation is usually steep, variations in its dip probably result from rheological layering of the crust and from asymmetric boudinage, an important and
ubiquitous process at various scales and development stages of the shear zone. That process eventually contributed to dismember the zone (Leloup et al., 1993).

The evidence for long-lasting high-temperature spells, with anatectic melting of crustal rocks and, apparently, some degree of mixing with rising mantle melts, is compelling. In the northern Ailao Shan, the temperature was high enough to cause unabated melting for at least \( \approx 4 \) Myr (26.3–22.4 Ma), while it remained above the solidus for nearly 2 Myr (24.2–22.4 Ma) in the Diancang Shan (Schärer et al., 1994). Such steady, high temperatures are consistent with shear heating at a sustained slip-rate of several centimeters per year (Nicolas et al., 1977; Fleitout and Froidevaux, 1980), under a depth-averaged shear stress on order of 50–100 MPa (Scholz et al., 1979), with none of the ambiguities that make the evidence from other shear zones, particularly in thrust terranes, inconclusive (e.g. Scholz, 1990).

The synoptic model that emerges (Fig. 33) is of a mature shear zone, comparable to that discussed, for instance, by Scholz (1990, fig. 3.26), but with magmatic melts contributing to weakening and strain-
Fig 32 (continued).
softening at and below the crustal brittle--ductile transition, and with downward extension into the mantle. The particularly high temperatures at relatively shallow depth, higher than suggested by simple shear-heating models (Fig. 33; Leloup and Kienast, 1993; Leloup et al., 1995), require upward transport of heat by magmatic fluids. The length and slenderness of the ASRR zone, the amount of displacement, discussed later, the location and alkaline chemistry of the Jianchuan monzo-syenites, which imply involvement of mantle magma sources (Schärer et al., 1994), require localized shear in the lithospheric mantle, roughly beneath that in the crust, as documented, for instance, along the Precambrian Norde--Stromfjord shear zone in Greenland (Sorensen, 1983). Finally, the fact that hypabyssal rocks intrude the shallow red beds of the Madeng and Midu gaps (Fig. 3) suggests volcanism along the zone in the Eocene, a circumstance impossible to establish in more ancient, deeply eroded terranes. Hence, as shown on Fig. 33, the ASRR shear zone may have acted for significant periods of time as a more or less continuous melt conduit, or cleft, all the way from the asthenosphere to the surface, which might ultimately account for its long lifespan as a weak deformation zone. The common occurrence of such a process within the continental lithosphere would help explain why models of deformation based upon bulk, solid-state flow-laws of rocks (e.g. England and Houseman, 1986) yield results that bear such little resemblance with the large-scale features observed in the India--Asia collision realm.

4.2. Concurrent evidence for a finite left-lateral offset > 500 km across the ASRR zone

4.2.1. Right- and left-lateral offsets

The finite, geological offset across the Red River Fault zone has long been a subject of debate. Confusion has arisen in part because motion along the zone switched from left-lateral in the Oligo-Miocene to right-lateral in the Plio-Quaternary.

Estimates have ranged from \( \approx 200--250 \) km right-lateral to more than 1500 km left-lateral. Before starting a critical discussion of the data at hand, let us recall that, given plate tectonics and scaling laws, it would be foolish to search for finite displacements smaller than a few tens of kilometers on a strike-slip shear zone 1000 km long. Nor should there be any hope to map such displacements by studying the zone in detail over a length of less than \( \approx 100 \) km.

The idea that the fault zone has offset dextrally the Yangbi and Chuxiong basins (Figs. 2 and 34b) by \( \approx 200--250 \) km (e.g. Cobbold and Davy, 1988; Dewey et al., 1989), a speculation based on the apparent, present-day separation of their western edges on 1:5,000,000 scale maps, can be dismissed at once. For such a finite geological offset to subsist, right-lateral strain would have had to overwhelm left-lateral strain along the zone, which is clearly not the case, the latter dwarfing the former by at least one order of magnitude (e.g. Allen et al., 1984; Lacassin et al., 1993b). That inference is all the more unfounded that the Chuxiong basin is just one of several upper Mesozoic basins north of the fault zone (Figs. 1, 2 and 34b).

In fact, all the prominent dextral offsets visible along the fault are of geomorphic, not geologic, nature. The best documented offset (\( \approx 6 \) km) is that of the Red River southeast of Daqiao (Fig. 35) and northwest of Atu, and of many of its tributaries in between (Allen et al., 1984). This offset, which has been estimated, from inferences on incision and regional uplift rates, to have accrued since sometime between the mid-Pleistocene (750 ka) and the Late Pliocene (3 Ma), has been used to place bounds on the current dextral slip-rate (2--8 mm/yr; Allen et al., 1984). It probably falls short of the total dextral offset on the fault, however. Larger valley offsets that help account for the complex geometry of the Red River's northern catchment where it crosses the Midu gap, can be found on satellite images and topographic maps. The clearest offset, about 20 km (Fig. 35), is between the deeply incised, \( \approx \) NE-trending Shuitian valley west of the fault (e.g. Allen et al., 1984, fig. 8), possibly a former, abandoned course of the Red River, and its entrenched, NE-trending channel east of the fault, past the bend downstream from Daqiao. Greater offsets (about 43 and 57 km, respectively) are obtained by matching the Daqiao or Shuitian valleys, west of the fault, with the Red River east of the fault and upstream from Danuo, where it tangents the Ailao Shan range-front (Figs. 3 and 35; Allen et al., 1984).
Fig. 33. Sketch of continental strike-slip fault at lithospheric scale. In case of Ailao Shan shear zone, NE is to the right (Yangzi block) and SW to the left (Indochina). Indochina moves fast relative to asthenospheric mantle while Yangzi block is nearly fixed. (a) Section of shear zone through crust and lithospheric mantle. Because of shear heating, 300°C, 500°C and 700°C isotherms (upper limits of greenschist facies, amphibolite facies and hydrous partial melting, respectively) are shallower in shear zone. Shear heating in upper mantle may be strong enough to initiate partial melting of lower crust (Fleitout and Froidevaux, 1980), hence induce ascent of crustal melts in shear zone (Leloup and Kienast, 1993). Mantle-derived partial melts may also rise within shear zone. (b) Strength profiles in shear zone for: no shear heating (dashed line), shear heating and heat transport only by conduction (gray line), and shear heating with partial melting and heat advection within shear zone (black line). (c) Corresponding estimates of surface heat flux across shear zone. (d) Detail of shear-zone crustal structure. D.L. = potential décollement levels on either side of shear zone.
That the uplifted Ailao Shan metamorphics form a barrier generally impassable to drainage, guiding much of the Red River course, accounts for the fact that few offsets larger than 6 km have been found, and so far mostly near the Midu gap. Although the evidence for such offsets is limited, we infer 20–50 km to be a plausible range of values for the finite late Cenozoic dextral displacement on the active Red River Fault (Fig. 36). If such amounts of displacement had accrued at a uniform slip rate since \( \approx 5 \) Ma, consistent with the onset of the second uplift phase in the Diancang Shan, that rate would be on the order of \( 7 \pm 3 \) mm/yr, slightly greater than that \( (5 \pm 3 \) mm/yr) inferred by Allen et al. (1984).

There is, by contrast, no shortage of convincing clues for very large cumulative sinistral offsets on the ASRR shear zone. When taken together, several lines of independent evidence point to the inescapable conclusion that the finite amount of left-lateral displacement taken up by that zone is in excess of 500 km, even though quantitative data are presently insufficient to constrain the exact value.

4.2.2. Strain rate and finite strain in the gneisses

A shear strain rate of 2–3 cm/yr for only \( \approx 4 \) Ma, probably a lower bound to account for a continuous melting span of that duration in the northern Ailao Shan (e.g. Scholz, 1990), would require a finite left-lateral offset on order of 100 km. If maintained for what we infer to be the likely minimum duration of left-lateral shear (\( \approx 18 \) Ma), such a rate would have produced a finite geological offset of 360–540 km (Fig. 36).
An offset of that size is consistent with the strain features observed and, where feasible, measured in the Ailao and Diancang Shan gneiss cores. The pervasive compositional banding, the overall mylonitic strain style, the ubiquitous occurrence of asymmetric foliation boudinage, and the frequent presence of sheath folds (Figs. 12–17, 20, 21, 31 and 32), all qualitatively indicate shear strains ($\gamma$) in excess of 20 (e.g. Bell and Etheridge, 1973; Quinquis et al., 1978; Lacassin and Mattauer, 1985; Gaudemer and Tapponnier, 1987; Malavieille, 1987; Lacassin, 1988). Quantitative support for this inference is provided by measurements of the layer-parallel extension of stretched amphibolites within the paragneisses. Surface-balanced restoration of several foliation-parallel amphibolite boudins trails at one site in the northern Ailao Shan yields $\gamma = 33 \pm 6$ (Lacassin et al., 1993b). Portions of the shear zone are less deformed. This is mostly the case of anatectic melts and intrusives, which are generally late-comers within various parts of that zone. Accordingly, the same restoration technique, applied to stretched melt veins at two distant sites along the Ailao and Diancang Shan, yields $\gamma = 7 \pm 3$. On the other hand, tens of meters wide ultramylonitic corridors, in which strain is beyond measurement, commonly occur in the gneisses. In addition, as shear went on for millions of years, progressive strain kept obliterating many anterior deformation features, so that those now visible in the rocks testify to only a fraction of the bulk finite strain. This suggests that the finite deformation observed across the 10 km wide ASRR zone is compatible with a minimum amount of left-lateral displacement of perhaps 200 km, if allowance is made for minor departures from a simple shear regime, or of $330 \pm 60$ km, if the strain measured in the amphibolites is taken to be a typical average value (Lacassin et al., 1993b).

4.2.3. Geological offsets

Given the nature and strain style of the ASRR shear zone, examination of the way in which it relates to the main features of the regional geology and tectonics, discussed in the first part of this paper, brings forth first-order constraints on its finite left-lateral offset (Figs. 1, 2, 3, 6 and 34). That there is no geochronological record of major tectono-metamorphic events along the zone between the late Proterozoic and the Tertiary simplifies the task of looking for offsets since, save for differential strain within regions on either side, one should expect most transverse Phanerozoic markers to be offset by roughly the same amounts.

To begin, recall that the Late Eocene–mid-Miocene sinistral motion recorded in the high-grade gneisses of the ASRR zone is in full kinematic agreement with the main phase of regional shortening to the south and north, a compatibility in large-scale strain that we have long pointed out (e.g. Tapponnier et al., 1986, 1990; Peltzer and Tapponnier, 1988; Leloup, 1991). The maps and sections of Figs. 3, 4, 6–12, 28 and 30 document the style of that phase, which is shown to have prominently affected the Late Cretaceous–Paleogene red beds and more ancient rocks over a widespread area. The manner in which the deformation of those rocks south of the zone blends with shear within it brings further support to an intimate transpressive link. The cause of such shortening and transpression in the Eocene–Oligocene is unambiguous: there exists no other event to invoke than the collision between India and Asia.

Three main types of geological structures north and south of the ASRR zone may be used to estimate finite sinistral offsets: upper Mesozoic red-bed "basins", upper Paleozoic–lower Mesozoic "sutures", and belts or outcrops of intrusive or extrusive rocks in that same age range (Fig. 34). As is commonly the case, there are problems with relying upon any such category of structures and rocks as large-scale displacement markers. First, while deeply underthrust tectonic imbricates, as in a suture zone, or intrusive batholiths, as in a magmatic arc, may be considered rather perennial transcrustal markers, effusive volcanics or sedimentary deposits are little more than surface coatings, the initial attitudes of which are easily modified by shallow-crustal processes.

One problem with the red-bed "basins", for example, is knowing whether they actually represent localized zones of subsidence and deposition or, instead, vestigial patches of a cover of formerly greater extent. The typical thickness (4–5 km) of the sequences in several such "basins" implies that the former is usually the case (e.g. Fig. 4), but the location of the original depocentres is often impre-
cisely known. Whether the present limits of the basins coincide with their initial depositional boundaries, are of a tectonic or erosional nature, or have significantly changed shape since deposition, raises additional problems. The western boundaries of the Chuxiong, Yangbi (or Lamping) and Khorat basins, for instance, are clearly tectonic. Such is the case, also, of all the boundaries of the Simao, Dongba, Qamdo–Markham, and Sichuan basins. The eastern limit of the Chuxiong basin and the southeastern limit of the Khorat basin, on the other hand, appear to be mostly depositional/erosional. In the Sichuan basin, the greatest thickness of red beds (≈ 10 km) lies in the east, along the Lungmen Shan, implying a Mesozoic flexural origin. The majority of the red-bed "patches" of eastern Tibet, South China and Indochina thus appear to be both basins, mostly of foreland type, and large synclinoria of Mesozoic–early Tertiary age. If, as our study of the regional tectonics suggests, the principal factor that changed the shape of those basin boundaries is the mostly Eocene (≈ 55–30 Ma) folding and thrusting caused by the early impingement of India into Asia, then they may safely be used as first-order markers of Oligo-Miocene offsets.

Other difficulties arise from the lack of continuity of marker-features at the scale involved. Constraining offsets on a fault zone with piercing points requires that fairly narrow, markedly oblique, and well-defined features be followed to near intersection with that zone. Given the geologic history and the accuracy of extant mapping, this condition is rarely met. The sutures and intrusive belts, potentially the best markers, crop out in discontinuous fashion because the younger red beds tend to cover them. In addition, because Tertiary strain caused tectonic uplift along the ASRR zone and involved left, or right-lateral motion on faults parallel, or at high angle to it, respectively, neither such markers nor all the boundaries of the red beds basins are easily traced to the very edges of that zone.

Once such difficulties are surmounted, wherever possible, two more sources of uncertainty occur. The first, a common pitfall, comes from the possible occurrence of several features of the same type on either side of the zone. This typically arises if the marker rocks are common, which is obviously the case here with the Mesozoic red beds and Permian basalts (Fig. 34b,c), but also with the Phanerozoic sutures (Fig. 34a), because there are several of them and they are generally poorly dated. For instance, potential counterparts of the Song Ma suture of Tongking (Figs. 1 and 2), whose age is still contentious, are unclear northwest of the ASRR zone. The second source of uncertainty is due to smearing along the shear zone, which results in smaller apparent offsets. That such a process took place is implied by the presence of low-grade schists and ultramafics along the southern Ailao Shan. In all likelihood, this process also affected the Permian basalts and Triassic andesitic–dacitic tuffs between Deqen and Luchun, as well as rarer rocks found only near the shear zone.

Given all the possible sources of error and uncertainty discussed, and in spite of the correspondingly large error bars and dispersion, the sinistral offsets depicted in Fig. 34 and summarized in Fig. 36 and Table 1 show a fair degree of consistency.

The minimum left-lateral separation of three different rock types that have been sinistrally smeared along the Ailao Shan–Red River zone indicates that the lowermost bound of such offsets is greater than 400 km. First, the belt of intermediate Triassic tufts and agglomerates that terminates near the Jianchuan–Madeng section north of the zone, only resumes south of it, west of the Yuangjiang–Mojiang road, at a distance greater than 400 km (Bureau of Geology and Mineral Resources of Yunnan, 1983) (Figs. 3 and 34c). Second, the thick Permian basalts last exposed between Midu and Lijiang north of the zone, are not found again south of it before Jlinping near the Vietnam border, at a distance of 460 km (Leloup, 1991; Wu Genyao, 1993) (Figs. 3, 6 and 34c). The typical, mostly alkalic petrology and chemistry of those rift basalts, whose pillows imply shallow-marine emplacement, is identical (Wu Genyao, 1993). Third, the only equivalent of the northern Diancang Shan and Jianchuan norites (N3) (Fig. 18), a distinctive, rarer rock type, is found at the extremity of the Ailao Shan near the Vietnam border, 500 ± 50 km to the southeast (Bureau of Geology and Mineral Resources of Yunnan, 1983; Leloup, 1991) (Fig. 34d).

Because outcrops of norites exist south of the Phang Si Pang, and both the tufts and basalts continue along one side or the other of the ASRR zone.
the actual offset of these three rock formations is probably larger. The total offset of the Triassic tuffs and arc volcanics between Deqen and the Dien Bien Phu fault, for instance, more likely ranges between 650 and 780 km (Figs. 2 and 34c). More Permian basalts are found near Benzilan (Fig. 6). The offset between these basalts and those at the Yunnan–Vietnam border would amount to 580 km or more (Fig. 34c). The other prominent patch of Permian basalts that extends south of the Sichuan basin to Chengjiang, southeast of Kunming, on the other hand, stops too far from the Red River to be used as a reliable marker, even though a match with basalts of that age in southern Tongking would yield comparably large numbers.

Matching the Benzilan–Jinsha suture of eastern Tibet with the Nan–Uttaradit suture of Thailand, inferring that the latter continues into Laos to Luang–Prabang, then along the dextral Dien Bien Phu fault, would yield an offset of 650–780 km (Fig. 34a), with smearing along the ASRR zone accounting for this range of values, as for the Triassic arc volcanics. Although such values are somewhat greater than those inferred by extending the northern end of the Nan–Uttaradit suture under the Simao basin (e.g. Peltzer and Tapponnier, 1988; Briais et al., 1993), the fact that we found clear evidence of dextral motion along the Benzilan suture may be taken as further justification for that match. Another possible match, and the same offset range, would be obtained by assuming that the Benzilan–Jinsha ultramafics mark the continuation of the Song Ma suture, first offset dextrally 100–200 km (Peltzer and Tapponnier, 1988) by the Dien Bien Phu fault. This alternative match would bring some elements of the late Paleozoic–early Mesozoic paleogeography into an acceptable framework, but would require the Jinsha and Song Ma sutures to have the same age, which remains to be established. Even if the Song Ma ultramafics were ultimately found equivalent to those reported in the Litang area (e.g. Mattauer et al., 1992) (Fig. 2), a sinistral offset of \( \approx 500 \) km on the ASRR shear zone would still be needed.

The most characteristic and continuous belts of intrusive rocks trending transverse to the Ailao Shan–Red River zone are the Permo-Triassic, Thai-Lincang range (granitic and granodioritic batholiths), and the Cretaceous granites that follow the coasts of South China, in Guangdong province, and of South Vietnam, between Vung Tau and Nha

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**Table 1**

Left-lateral offsets of geological markers across the Ailao Shan–Red River shear zone

<table>
<thead>
<tr>
<th>South China</th>
<th>Indochina</th>
<th>Left-lateral offset along the ASRR zone (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Ophiolites</strong></td>
<td>Jinsha–Benzilan</td>
<td>Song Ma or Uttaradit</td>
</tr>
<tr>
<td><strong>Mesozoic red-bed basins</strong></td>
<td>West Markam</td>
<td>West Yangbi</td>
</tr>
<tr>
<td></td>
<td>Dongba</td>
<td>West Yangbi</td>
</tr>
<tr>
<td></td>
<td>East Markam</td>
<td>East Simao</td>
</tr>
<tr>
<td></td>
<td>West Sichuan</td>
<td>West Khorat</td>
</tr>
<tr>
<td></td>
<td>East Chuxiong</td>
<td>East Khorat</td>
</tr>
<tr>
<td><strong>Upper Permian basalts (P(_2^2))</strong></td>
<td>Near Jianchuan</td>
<td>South of Yuanyang</td>
</tr>
<tr>
<td></td>
<td>Near Benzilan (section U1)</td>
<td>South of Yuanyang</td>
</tr>
<tr>
<td><strong>Triassic arc volcanics (T(_2)–(_3))</strong></td>
<td>West of Jianchuan (section S)</td>
<td>Southeast of Yuanjiang</td>
</tr>
<tr>
<td></td>
<td>Northwest of Weixi</td>
<td>Dien Bien Phu</td>
</tr>
<tr>
<td><strong>Cretaceous granites</strong></td>
<td>Western boundary</td>
<td>Western boundary</td>
</tr>
<tr>
<td><strong>Norites (N(_1^2))</strong></td>
<td>North of Diancang Shan</td>
<td>Near Jinping</td>
</tr>
</tbody>
</table>
Trang (Fig. 1). Southwest of the Red River, the Lincang batholith is both offset by the Nan Ting fault and interrupted by a narrow zone of high-grade mylonitic rocks (Chongshan metamorphic belt, Fig. 3) (Wu Haoruo et al., 1994), parallel to the Ailao Shan, that may have been a similar, Tertiary sinistral shear zone. This precludes the use of that batholith as a marker for offsets along the ASRR zone (Figs. 1, 2 and 34d). The granite belts of coastal Guangdong and South Vietnam, on the other hand, can be traced to within ≈ 150 and 50 km of that zone and of its extension offshore, respectively (Figs. 1 and 34d), without the occurrence of any large intervening tectonic feature. The origin of those granites, presently under study, is contentious, but the regional stratigraphy leaves little doubt on their Cretaceous age. Although the southernmost part of the Guangdong granite belt appears to have been rifted away from the mainland in the early stages of opening of the South China sea (e.g. Briais et al., 1993), matching the northwestern edges of the intrusive belts on either side of the ASRR shear zone yields an offset of 700 ± 100 km (Fig. 34d).

The most prominent red-bed basins of western Yunnan and eastern Tibet are abruptly truncated by the Ailao Shan–Red River shear zone, and thus have boundaries that define acceptable piercing points. On the northeast side of the zone, the closest possible counterparts of the Yangbi and Simao basins are the Dongba and Qamdo–Markham basins (Figs. 1 and 2). The Yangbi and Dongba basins contain similar sequences of Cretaceous–Eocene limestones and red sandstones with comparable styles of Tertiary folding. The separation between the piercing points of the western boundaries of the Markham and Yangbi basins is 500 ± 50 km (Fig. 34b). This offset value is the smallest for this group of basins. Matching the eastern boundaries of the Simao and Markham basins, for instance, would yield an offset of 700 ± 50 km. Matching the western boundary of the Yangbi basin with that, less well defined, of the Dongba basin, would yield an offset in excess of 600 km (Fig. 34b).

Even greater offsets are suggested by the separation of the largest red-bed basins of Southeast Asia, farther away from the ASRR zone. Those offsets are more uncertain because the corresponding markers, except one, cannot be simply traced to the vicinity of the shear zone. Also, such offsets probably integrate significant finite sinistral displacements on other Tertiary faults, the most prominent of which being the Xianshuihe, Song Da–Song Ma, and Song Ca fault zones (Fig. 2). With this in mind, a displacement of 800 ± 200 km, for instance, is needed to realign the western, tectonic boundaries of the Sichuan and Khorat basins (Figs. 2 and 34b), as long advocated by some of us (Tapponnier et al., 1986; Peltzer and Tapponnier, 1988). Matching the eastern, depositional/erosional boundaries of the Chuxiong and Khorat basins would imply as much as 1050 ± 100 km of sinistral displacement (Fig. 34b).

The geological evidence at hand (Table 1 and Fig. 36) is thus consistent with a minimum Tertiary left-lateral offset on the Ailao Shan–Red River shear zone (sensu stricto) greater than 400 km. On a broader scale, if other Cenozoic sinistral faults roughly parallel to this shear zone are included, the maximum amount of Tertiary displacement between the stable parts of the Yangzi and Khorat–Kontum platforms might have reached about 1200 km. At a more detailed level, the evidence discussed implies a minimum of ≈ 500 km and a most likely value of 700 ± 200 km for the Tertiary sinistral offset on the ASRR zone per se (Table 1 and Fig. 36). Note that the Plio-Quaternary dextral offset on the Red River fault, if on order of several ten kilometers, would substantially increase the latter values. Given the extant geochronological constraints, which imply that the shear zone was active for about 20 Myr, a finite displacement of about 700 km would correspond to a uniform shear-strain rate of 3.5 cm/yr. Such a rate would be compatible with the mechanical inferences made earlier.

4.2.4. Paleomagnetic evidence and seafloor spreading

Systematic sampling of red beds in the basins of South China and Indochina, still in progress, has begun to yield a fairly dense upper Mesozoic paleomagnetic database, with which assessment of large-scale latitudinal Tertiary motions and rotations has become feasible. Several different studies have shown, for instance, that South China as a whole has not rotated significantly since the Cretaceous (e.g. Enkin et al., 1991, 1992; Gilder et al., 1993; Huang and Opdyke, 1993; Ma et al., 1993). Since that time, on the other hand, the Khorat red beds, on the stable
core of Indochina, appear to have experienced 14.2 ± 7.1° of clockwise rotation relative to those of Sichuan in South China (Yang and Besse, 1993). Such studies thus imply significant sinistral motion of Indochina with respect to a relatively stable South China block. From the differences between observed and predicted Cretaceous paleolatitudes, Huang and Opdyke (1993) have inferred several degrees of southward translation of the Simao basin with respect to stable Eurasia, an amount unfortunately within measurement uncertainties. Using an improved, more complete database, Yang and Besse (1993) and Yang et al. (1995) have concluded that the post-Cretaceous southward motion of Indochina relative to China amounted to 7.9 ± 2.5° in latitude, which may be taken as evidence for as much as 1200 ± 500 km of left-lateral displacement on the Red River fault system (Fig. 36). While this particularly large value is compatible, within uncertainties, with the offset we find most likely on the ASRR shear zone (700 ± 200 km), we suspect that, like the largest geological offsets listed in Table 1 and for the same reasons, it integrates finite sinistral displacements on other faults roughly parallel to that zone (Figs. 2 and 34) (Yang and Besse, 1993).

One final, independent way to constrain the amount of offset on the ASRR shear zone, quantitatively pioneered by Briais et al. (1993), is based on the idea that seafloor-spreading in the South China sea, which the Ailao Shan–Red River shear zone enters south of Hanoi, has absorbed most of the left-lateral movement along that zone (e.g. Tapponnier et al., 1982, 1986; Peltzer and Tapponnier, 1988). Particularly strong support for this idea comes.

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**Fig. 36.** Summary of geomorphic (right-lateral) and geologic (left-lateral) offsets across the Ailao Shan–Red River shear zone.
from the new radiometric ages found in the ASRR metamorphics, which establish that the duration of shear on the zone (35–17 Ma; Leloup et al., 1993; Schärer et al., 1994; Harrison et al., 1995) was nearly coeval with that of spreading in the sea (32–15.5 Ma; Briais et al., 1993). By simply assuming the shear zone and the spreading ridge to form a single boundary between two plates (Indochina and South China), the rotation parameters derived from the magnetic anomalies on the seafloor can be used to reconstruct motion along that zone (Briais et al., 1993). The reconstruction is consistent with the ASRR shear zone being mostly strike-slip, with transpression in the northwest and transtension in the southeast (Briais et al., 1993, fig. 14). For a point situated between Yuanjiang and Yuanyang (Fig. 3), the sinistral shear-strain rate is found to range between 3.0 and 3.7 cm/yr, and the finite amount of left-lateral movement is $\approx 540 \pm 50$ km (Briais et al., 1993) (Fig. 36). This latter value must be considered a lower bound since it relates to seafloor-spreading only. Including the strain related to rifting and crustal attenuation on either margin of the sea prior to spreading would plausibly increase the amount of movement on the ASRR shear zone by 100–200 km, assuming the mechanism to have remained the same. Hence, both the rate and the finite sinistral offset derived from seafloor-spreading parameters appear to be in good agreement with those inferred from geological evidence onland.

4.3. Role of the ASRR shear zone in the tectonics of India–Asia collision

Overall, the observations and results presented and discussed in this paper thus strengthen the long advocated view that the Ailao Shan–Red River shear zone is the greatest Cenozoic tectonic feature of continental Southeast Asia east of the Assam syn-taxis. That it absorbed many hundreds of kilometers of sinistral motion in the Oligo-Miocene, as Indochina was extruded southeastwards, relative to a then more stationary South China, by the penetration of India into Asia, is inescapable. Even though it slices deep into the interior of a continent, the ASRR zone has in fact most of the attributes of a great transform plate boundary. It is thus no surprise that it played a key role in orchestrating the tectonics of a vast region and the rearrangement of large lithospheric blocks, all the way from Tibet to Borneo.

A measure of the importance of that rearrangement, and of its bearing on collision mechanics, is readily obtained by estimating the surface of Asian lithosphere removed by such transform faulting from regions facing the Indian indenter (e.g. Tapponnier et al., 1986; Peltzer and Tapponnier, 1988; Armijo et al., 1989). To a first order, that surface loss (extrusion) may be derived from the quantitative reconstruction of Briais et al. (1993), with a couple of amendments. First, we adjust Briais et al.'s finite rotation parameters — pole at 7.9°N, 85.7°E, $\omega = 10.8^\circ$, from anomaly 10 (30.3 Ma) only — to include anomaly 11 (32 Ma) and to allow for $\approx 150$ km of crustal extension across the South China Sea margins prior to seafloor spreading. The slightly different parameters we obtain (pole at 8.2°N, 85.6°E, $\omega = 15.2^\circ$) (Fig. 37) hence fall in better agreement with 700 km of sinistral motion along the ASRR zone and with the mean paleomagnetic rotation ($14.2^\circ$) between the Sichuan and Khorat red beds. Second, we take 150–200 km of Eocene sinistral slip on the Wang Chao–Three Pagodas Fault zone (Figs. 1 and 37) (Peltzer and Tapponnier, 1988; Lacassin et al., 1993b, in prep.) into account by assuming such motion to correspond to 6.5° of additional clockwise rotation of South Indochina about the adjusted Indochina/South China pole (Fig. 37). The surface of the roughly triangular zone between the current western boundary of Indochina and its position prior to rotation, $\approx 40$ Ma (shaded on Fig. 37), is then computed on the sphere. The area thus lost to extrusion amounts to about $830,000$ km$^2$ with $\approx 640,000$ and $\approx 190,000$ km$^2$ due to movement along the Ailao Shan–Red River and Wang Chao–Three Pagodas zones, respectively. Prominent field evidence for shortening and thickening of the crust of that triangular region, which lies in the path of India’s eastern tip, both during and after extrusion (Tapponnier et al., 1990; Lacassin et al., 1995), implies that this amount must be considered a lower bound. Using stage rotation poles located farther southwest than on Fig. 37 (Briais et al., 1993) would also increase the extruded area. Hence, the comparison of the area obtained here ($0.83 \times 10^6$ km$^2$) to various estimates of the total Asian surface loss due to collision — $6–8.5 \times 10^6$ km$^2$, according to Tapponnier et al.
(1986); \(3.5-4.2 \times 10^6 \) km\(^2\), according to Le Pichon et al. (1992) — requires that \(17 \pm 7\%\) of the penetration of India into Asia, at least, was absorbed by the extrusion of Indochina alone.

Much still remains to be elucidated. The significant differences, for instance, between the various offsets estimated from the regional geology and paleomagnetism need be resolved by finer-scale fieldwork in the mountainous countries that straddle the shear zone, many of which have remained until recently beyond reach. Much more detailed mapping, analysis, and radiometric dating of mafic–ultramafic and granitic rocks need be done in order to convincingly assess the exact correspondence between sutures and magmatic arcs that have been offset by the shear zone. An even more daunting task lies ahead to determine the precise offshore connection, outside the realm of hand-sampling and accurate satellite or aerial imaging, of the various strands of the Ailao Shan–Red River system with the extensional basins and spreading centers of the South China sea. The limited shallow seismic reflection data, mostly

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**Fig. 37.** Minimum estimate of extruded area. Present-day coastlines of South China, Indochina and India (continuous lines) are plotted together with their position before Indochina extrusion (= 40 Ma, dotted lines).
"classified", and the marine geophysical evidence acquired to date are clearly inadequate to address that facet of the problem.

At least does it seem safe to conclude, with the facts gathered on land at this point, that all interpretations or models of Southeast Asian tectonics that do not allow for, or predict, the existence of the Ailao Shan–Red River shear zone, with at least 500 km of Oligo-Miocene sinistral motion on it and, by way of consequence, an intimate causal link between that zone and the opening of the South China sea (e.g. Dewey et al., 1989; England and Houseman, 1986; Cobbold and Davy, 1988; Sornette et al., 1993; Rangin et al., 1995) ought to be regarded as fundamentally flawed, be it in terms of the geological or of the physical inferences upon which they rest.

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