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ABSTRACT: The *Tethysides* are a superorogenic complex flanking the Eurasian continent to the south and consisting of the *Cimmerides* and *Alpides*, products of *Palaeo-* and *Neo-Tethys* respectively. We here review their evolution, mainly on the basis of new maps showing the distribution of sutures, magmatic rocks, certain palaeobiogeographically and palaeoclimatologically significant taxa and facies, and fragments of Pan-African (900–450 Ma) orogenic system forming the basement of many Tethyside blocks. These are supplemented by palaeomagnetic data reported in the literature.

A fundamental tenet of this paper is that major sutures which contain ophiolite fragments, represent tectonic sections between continental blocks where oceanic crust has been subducted. Palaeo-Tethys came into existence largely in late Carboniferous time. Coevally, it began to be consumed by both internal and peripheral subduction zones, which continued into the Permian; some of these had been inherited from pre-Tethyan times. In the later Permian, rifting subparallel with the northern margin of Gondwana Land began between the Zagros and Malaysia, separating a Cimmerian continent from N. Gondwana Land, and thus heralding the opening of Neo-Tethys and other smaller oceans that were back-arc basins of Palaeo-Tethys. This rifting possibly also extended farther west into Crete and mainland Greece. However, the North China block, Yangtze block, Huanan block, the eastern moity of the Qangtang block (North Tibet), and Annamia, all originally pieces of the end-Proterozoic–early Palaeozoic Gondwana Land, had already separated from it in pre-late Carboniferous times, possibly during the Devonian. All of these blocks, and the Cimmerian continent, were characterized by Cathaysian floral elements in late Palaeozoic time. Palaeomagnetic and palaeontological data showing the original Gondwana Land affinity of these continental blocks are supplemented by correlating late Proterozoic–early Palaeozoic Pan-African sutures, orogenic belts, and sedimentary basin fragments across Tethyside sutures. Late Permian foraminiferal provinces are related to this palaeogeographical interpretation.

By Triassic times, most Cimmeride subduction zones were already in existence. The Cimmerian Continent accelerated its separation from Gondwana Land and—locally in the late Permian—began disintegrating internally along the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean. By late Triassic time all of the Chinese blocks—except Lhasa—and Annamia had collided with each other and with Laurasia. The resulting enormous orogenic collage had a 'soft cushion' between itself and Laurasia, in the form of the enormous accretionary complex of the Songpan–Ganzi. This connection enabled Laurasian land vertebrates to reach south-east Asia by late Triassic time. In late Triassic to middle Jurassic times, most major Cimmeride collisions were completed. Widespread aridity in Central Asia occurred in late Jurassic time, probably in the rain shadow of the newly formed Cimmeride mountain wall.

Neo-Tethyan subduction systems formed along the S. margin of the Cimmerides or within Neo-Tethyan oceanic lithosphere during the Jurassic. Most, if not all, were north- or east-dipping. They continued the northerly migration of the Tethyside blocks.

Evolution of the Tethysides influenced the distribution of marine and terrestrial organisms, and affected sea-level changes and patterns of atmospheric circulation during much of the Mesozoic and Cainozoic. It is likely to have reflected the surface expression of a persistent trend in the large-scale convective circulation in the mantle, that continuously transported material northward into the Tethyan domain.

Within the Alpine–Himalayan–Indonesian mountain ranges, two distinct but largely superimposed orogenic systems exist, parts of which correspond with Suess' (1909, 1911) original *Cimmerian Mountains* and the *Alpides* (Şengör 1984). Respecting Suess' priority, Şengör (1984) called the older of these two systems *Cimmerides*, and the younger *Alpides*, together forming the superorogenic complex *Tethysides*. Both of these systems have been shown to include a very large number of sutures, but two dominant times of ocean closure along

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them are discernible: the earlier interval from the late middle Triassic to earliest middle Jurassic, corresponds with the formation of the Cimmerides, whereas the later episode, the Alpide, was dominantly from the late Palaeocene to late Eocene. Although in both systems a number of suture segments closed outside the intervals indicated, these intervals appear valid for such long stretches that it may be possible to schematize them into a 'northern group' of Cimmeride sutures and a 'southern group' of Alpide sutures, with a long Cimmerian continent (or a continental archipelago), and a number of 'exotic' continental fragments in between (Fig. 1) (Şengör 1979*a*, 1984). This grossly simplified scheme makes a rapid overview easy, and also provides a practical nomenclature for large-scale palaeogeographical and palaeotectonic elements of Pangaea in the Old World. We have, thus, Laurasia in the north (Permian to Cretaceous), Palaeo-Tethys to its south (early Carboniferous to middle Jurassic), the Cimmerian continent in the middle (Triassic to middle Jurassic), Neo-Tethys between the Cimmerian continent and northern Gondwanaland (? latest Permian or Triassic to largely Eocene, locally still extant) and finally Gondwana Land itself in the south (*c.*Ordovician to Jurassic) (Fig. 1) (Şengör 1984, Şengör and Hsü 1984).

The closure of these two sets of 'Tethyan' sutures generated the Cimmeride and Alpide orogenic systems (Fig. 2), with their associated wide and complex areas of cratonic disruption that now dominate the architecture of nearly the whole of Eurasia. Palaeo- and Neo-Tethys, plus the continental portions separating them or contained within them may be designated as the *Tethyan domain* or the *Tethyan realm*.

Although the simple classification given in Fig. 1 serves as a useful terminological basis for further research, the suture map displayed in Fig. 3 emphasizes how much more complicated the real situation is. The Tethysides consist of an enormous orogenic collage, containing a large number of continental and otherwise buoyant blocks, stitched together by an anastomosing suture network. Following the dual division of the Tethysides, this *Tethyside collage* also con-

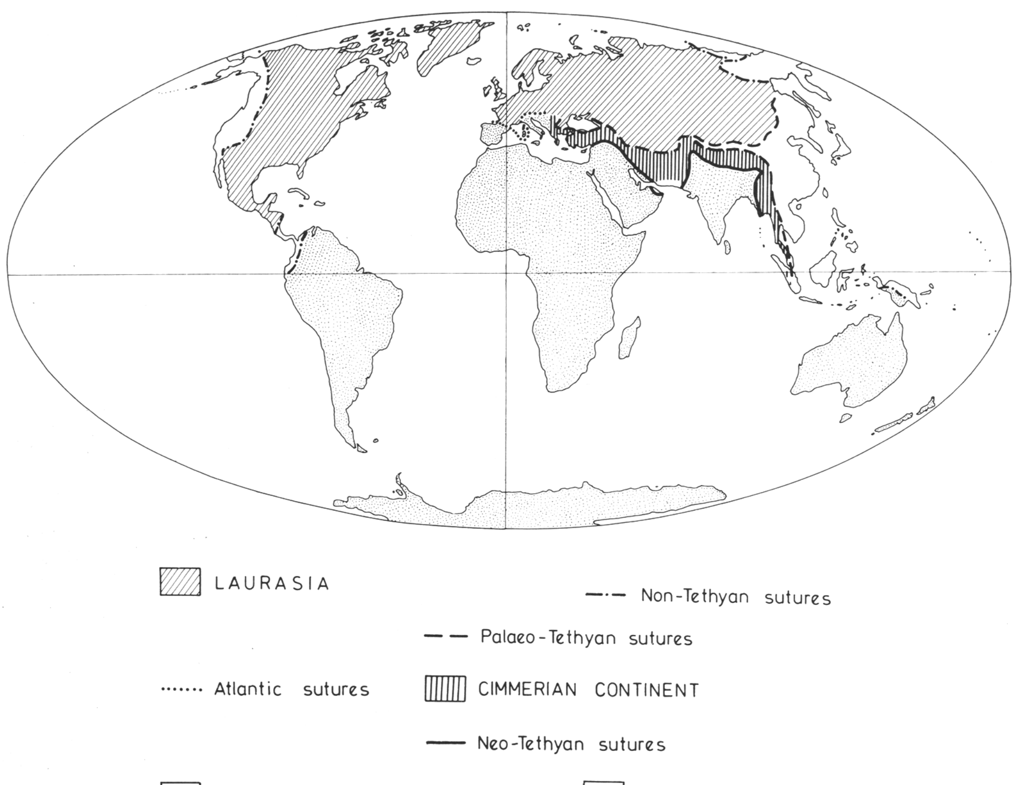


FIG. 1. First-order palaeogeographical/palaeotectonic elements taking part in the architecture of the Tethysides, and their place in the present structure of our planet.

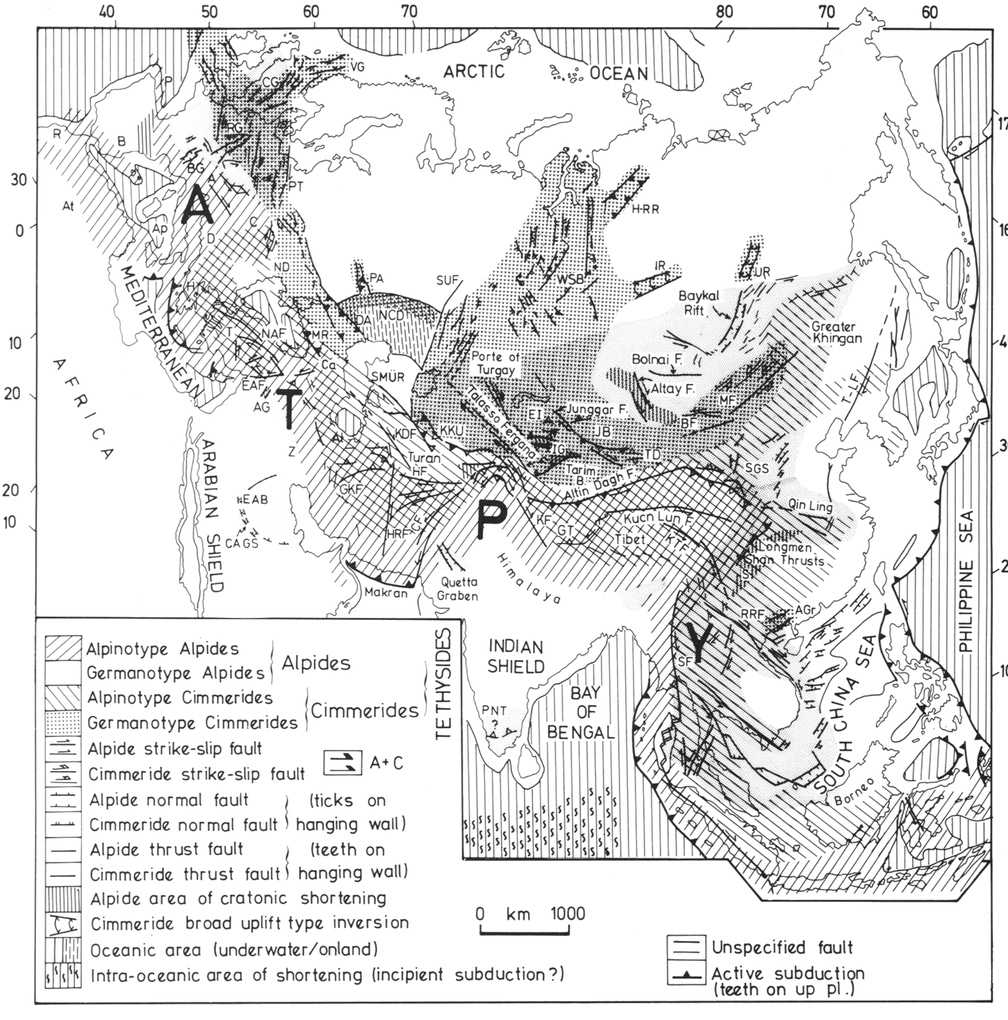


FIG. 2. The Tethyside superorogenic complex, showing both the Cimmerides and the Alpides and their associated areas of fore- and hinterland deformation. Notice the degree of overprinting of the Cimmerides by the Alpides, especially between the Carpathians and China, which has long hindered the recognition of the former as an independent orogenic belt. Key to lettering: larger bold letters *A*, *T*, *P*, and *Y* are the Alpine, Turkish, Pamir, and Yunnan syntaxes respectively. Smaller letters: *A*—Alps, *AG*—Akçakale Graben, *AGr*—An Chau Graben, *Al*—Alborz, *Ap*—Apennines, *At*—Atlas mountains (*sensu lato*), *B*—Betics, *BF*—Bogdo Fault, *BG*—Bresse Graben, *C*—Carpathians, *Ca*—Caucasus, *CAGS*—Central Arabian graben system, *CF*—Chaman Fault, *CG*—Central Graben, *D*—Dinarides, *DA*—Dnyepr–Donetz aulacogen, *EAB*—East Arabian block, *EAF*—East Anatolian Fault, *EI*—East Ili Basin, *GKF*—Great Kavir Fault, *GT*—Gerze thrust, *H*—Hellenides, *HF*—Herat Fault, *HRF*—Harirud Fault, *H-RR*—Hantaj-Rybninsk Rift, *IG*—Issyk Gol Basin, *IR*—Irkineev Rift, *KDF*—Kopet Dagh Fault, *KF*—Karakorum Fault, *KKU*—Kizil Kum uplift, *KTF*—Kang Ting Fault, *MF*—Mongolian Faults, *MR*—Main Range of the Greater Caucasus, *NAF*—North Anatolian Fault, *NCD*—North Caspian depression, *ND*—North Dobrudja, *PA*—Pachelma aulacogen, *PNT*—Palni–Nilgiri Hills Thrust, *PT*—Polish Trough, *RG*—Upper Rhine Graben, *RRF*—Red River Fault, *S*—Sichuan Basin, *SF*—Sagain Fault, *SGS*—Shanxi Graben System, *SMÜR*—South Mangyshlak–Üst Yurt Ridge, *SUF*—South Ural Faults, *T*—Turkish Ranges, *TD*—Turfan depression, *T-LF*—Tan-Lu Fault, *UR*—Ura Rift, *VG*—Viking Graben, *WSB*—West Siberian Basin, *Z*—Zagrides.

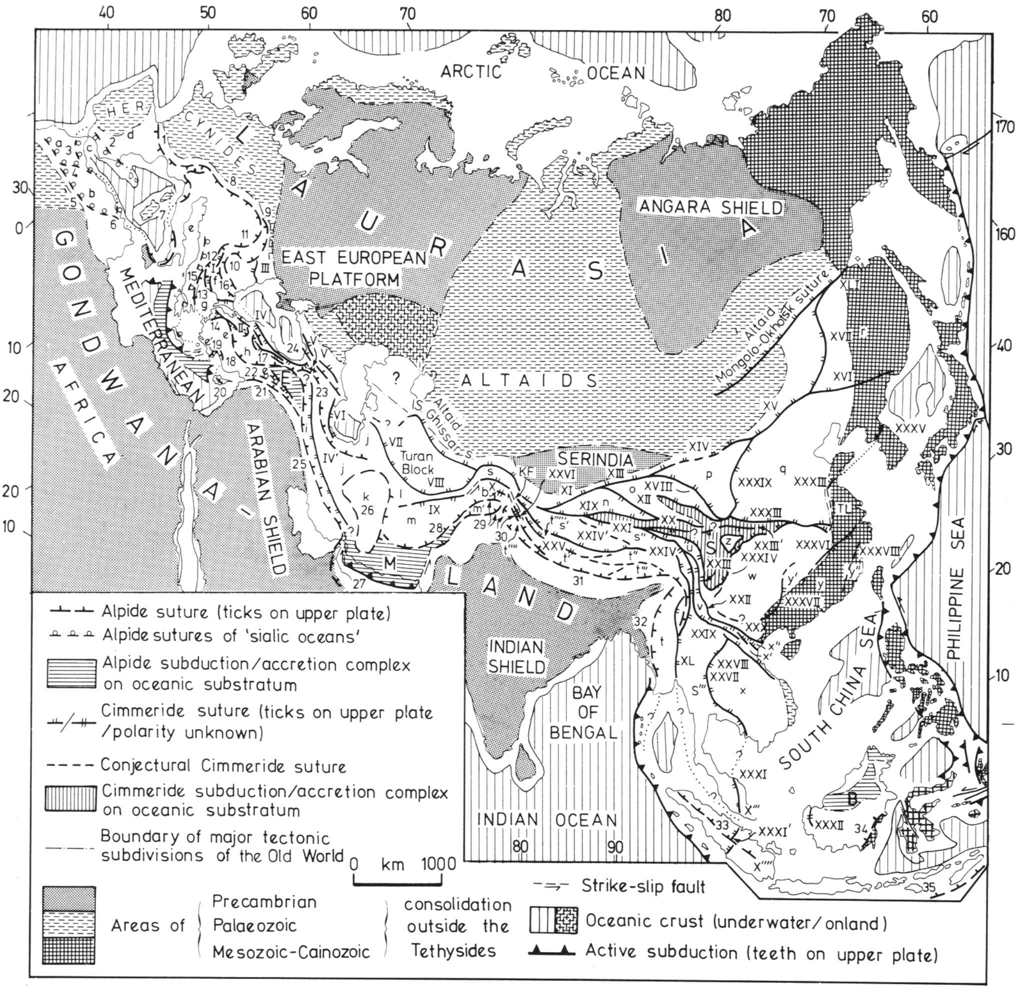


FIG. 3. Tectonic map of Eurasia and North Africa, showing their major tectonic subdivisions and the distribution of sutures within the Tethysides. Tethysides correspond with white areas on land. Cimmeride sutures: *I*—Palaeo-Tethyan suture in Greece and Yugoslavia, *II*—Karakaya, *III*—Luncavita–Consul, *IV*—North Turkish, *IV′*—Sanandaj–Sirjan suture, *V*—Svanetia, *V′*—Chorchana–Utslevi, *VI*—Talesh, *VII*—Kopet Dagh/Mashhad, *VIII*—Paropamisus/Hindu Kush/North Pamir, *IX*—Waser, *X*—Rushan–Pshart, *XI*—Northern Kuen-Lun, *XII*—Qimandag, *XIII*—Altin Dagh, *XIV*—Suelun–Hegen Mountains, *XV*—Inner Mongolian, *XVI*—Suolun–Xilamulun, *XVII*—Greater Khingan, *XVIII*—Tergun Daba Shan/Qinghai Nanshan, *XIX*—Southern Kuen-Lun, *XX*—Burhan Budai Shan/Anyemaqen Shan, *XXI*—Hoh Xil Shan/Jinsha Jiang, *XXII*—Maniganggo, *XXIII*—Litang, *XXIII′*—Luochou 'arc-trench belt', *XXIV*—Banggong Co–Nu Jiang, *XXIV′*—'Mid-Qangtang', *XXV*—Shiquan He, *XXVI*—south-west Karakorum, *XXVII*—Nan-Uttaradit/Sra-Kaeo, *XXVIII*—Tamky–Phueson, *XXIX*—Song Ma, *XXX*—Song Da, *XXXI*—Bentong–Raub, *XXXI′*—Mid-Sumatra, *XXXII*—Serabang (West Borneo), *XXXIII*—Qin-Ling/Dabie Shan, *XXXIII′*—northernmost Jiangsu, *XXXIV*—Longmen Shan/Qionglai Shan, *XXXV*—Chugareong, *XXXVI*—Shaoxing–Pingxiang, *XXXVII*—Tianyang, *XXXVIII*—Lishui–Haifeng, *XXXIX*—Helan Shan, *XL*—Mandalay, *XLI*—Shilka. Alpide sutures: *1*—Pyrenean, *2*—Betic, *3*—Riff, *4*—High Atlas, *5*—Saharan Atlas, *6*—Kabylian, *7*—Apennine, *8*—Alpine, *9*—Pieniny Klippen Belt, *10*—circum-Moesian, *11*—Mureş, *12*—Vardar, *13*—Peonias/Intra-Pontide, *14*—Almopias/İzmir–Ankara, *15*—Pindos–Budva–Bükk, *16*—Srednogorie, *17*—Ilgaz–Erzincan, *18*—Inner Tauride, *19*—Antalya, *20*—Cyprus, *21*—Bitlis (Miocene) and Assyrian (Senonian), *22*—Maden, *23*—Sevan/Akera/Qaradagh, *24*—Slate-Diabase zone, *25*—Zagros, *26*—circum-Central Iranian Microcontinent (including Sabzevar and Sistan), *27*—Oman, *28*—Waziristan, *29*—Kohistan (*29* North—Chalt, *29* South—Main

sists of two parts of unequal size. The older and much larger part is formed from fragments accreted to Laurasia during the Cimmeride evolution, and is called the *Cimmeride collage*. The younger and much smaller part was accreted to Eurasia during the Alpide evolution, and is called the *Alpide collage* (Şengör 1986*a*). In Fig. 3, the Cimmeride and Alpide sutures are identified with separate symbols to distinguish their components from one another.

In the following paragraphs we first comment on our improved suture map of the Tethysides (Fig. 3), and establish the timing of events along them. We then explore the provenance of the major components of the Cimmeride and Alpide collages, on the basis of palaeontological, palaeomagnetic, and other geological data, showing that most, if not all, of them originated in Gondwana Land. Following a highly schematized and tentative evolutionary history of the Tethysides, we point out the extremely important role of multiphase strike-slip motion in complicating the evolution of the Tethysides. We conclude by emphasizing the vast complexities and the tremendous mobility inherent in the evolution of orogenic complexes in general, and underline the misleading role of such simplistic methods as the 'terrane concept' that are employed for their analysis.

Suture zones of the Tethyside orogenic collage

Fig. 3 illustrates all of the known major suture zones within the Tethyside superorogenic complex. The most conspicuous feature of this map is the nearly complete independence of the Tethyside sutures from those of older orogenic complexes such as the Hercynides and the Altaids. Other than two exceptions (the Ghissar and the Mongolo–Okhotsk sutures: Fig. 3) none of the Tethyside sutures seem to connect with sutures outside the Tethyside suture network in Eurasia. One implication of this is that a few long and thin (present geometries) continental objects had collided with the older Altaid orogenic collage to 'seal it off' before the Tethyside collisions with Asia commenced. In the following sections we review first the Cimmeride, and then the Alpide sutures, mainly with respect to the ages of ocean opening and closing along them. Their orogenic polarities (i.e. pre-collision facing of main subduction zone) are shown in Fig. 3. Figs 4 through 7 display the evolution of magmatism along these suture zones, and provide a major basis for their identification.

Cimmeride sutures[1]

Palaeozoic Cimmeride sutures

A few of the earliest 'Tethyside' collisions had taken place between individual components of the Cimmeride collage (e.g. the pre- or intra-Devonian Tamky–Phueson suture in North Vietnam (*XXVIII*): Hutchison, in press, or the coeval Tergun Daba Shan–Qinghai Nanshan suture (*XVIII*): Yang *et al.* 1986) even before the triangular Pangaean gap came into existence. Thus, these sutures are not Tethyside in the strict sense. The reason why they are so considered here is that their closure seems to have triggered the onset of subduction activity along the nearby Tethyside sutures, thus exercising a

[1] In this paper Roman numerals in parentheses refer to Cimmeride sutures shown in Fig. 3, Arabic numerals refer to Alpide sutures, and lower case letters refer to Tethyside blocks. Hereinafter they are cited without repeated reference to Fig. 3.

Mantle Thrust), *30*—Ladakh (*30* North (N)—Shyok, *30* South (S)—Indus), *31*—Indus/Yarlung-Zangbo, *32*—Indo-Burman, *33*—Woyla, *34*—Meratus, *35*—Timor. Tethyside blocks: *a*—Moroccan Meseta, *b*—Oran Meseta, *c*—Alboran, *d*—Iberian, *e*—African Promontory (including the Menderes-Taurus block; *e'*—Alanya Massif), *f*—Rhodope-Pontide, *g*—Sakarya, *h*—Kırşehir, *i*—Sanandaj-Sirjan zone, *j*—north-west Iran, *k*—central Iranian microcontinent, *l*—Farah, *m*—Helmand, *m'*—Kohistan, *n*—western Kuen-Lun 'Central Meganticlinorium', *o*—Qaidam, *p*—Alxa (Ala Shan), *q*—North China ('Sino-Korean'), *r*—North China Foldbelt, *s*—Central Pamir-Qangtang-Sibumasu (*s'*—Central Pamir/West Qangtang, *s''*—East Qangtang, *s'''*—Sibumasu), *t*—Lhasa-Central Burma (*t'*—Bongthol Tangla, *t''*—Nagqu, *t'''*—Lhasa proper: A. Gansser, verbal comm. 1985, *t''''*—Ladakh, *t'''''*—west Lhasa-south Pamir), *u*—Shaluli Shan, *v*—Chola Shan, *w*—Yangtze, *x*—Annamia (possibly with two exotic island arcs of early (*x'*) and late (*x''*) Palaeozoic age and eastern Malay Peninsula (*x'''*) and south Sumatra (*x''''*)), *y* (*sensu lato*)—Huanan (*y* (*sensu stricto*)—Huanan, *y'*—Hunan, *y''*—coastal block), *z*—Songpan Massif. Key to lettering: *B*—Sarawak accretionary complex, *b*—South-east Pamir black slates, *E*—East Anatolian accretionary complex, *KF*—Karakorum Fault, *M*—Makran accretionary complex, *TL*—Tan-Lu fault.

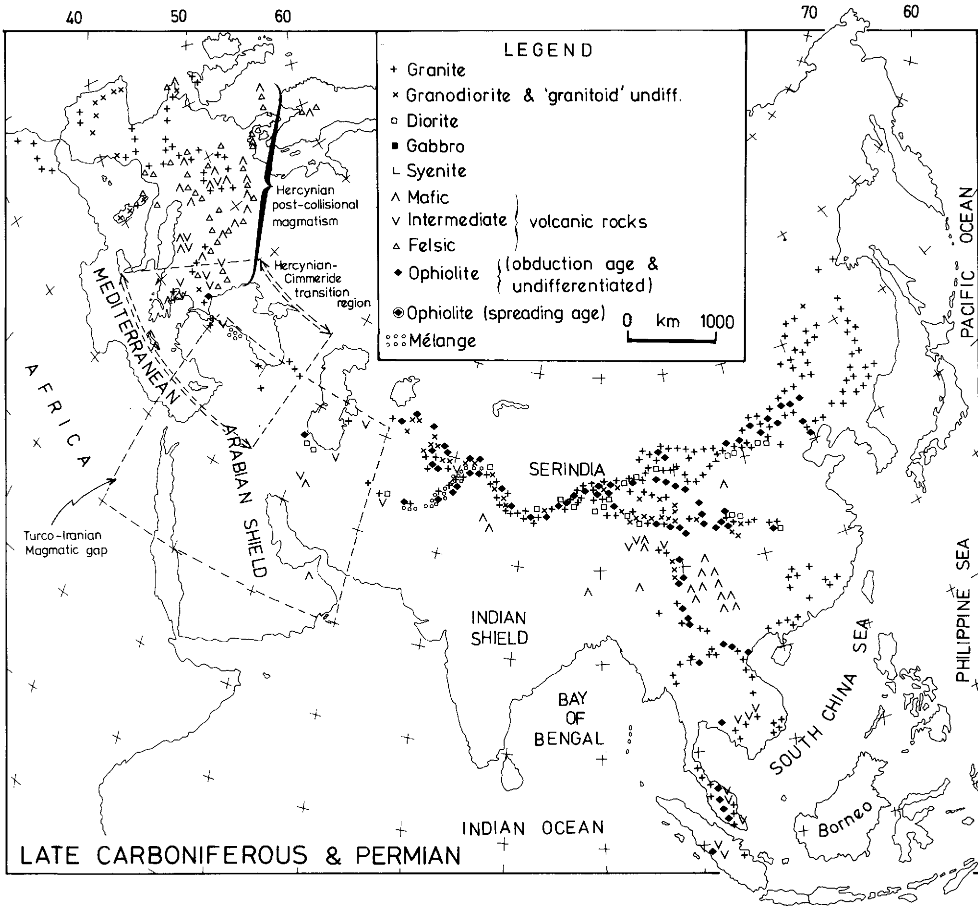


FIG. 4. Late Carboniferous and Permian magmatism in the Tethysides and in the Hercynides. Notice the conspicuous contrast between the irregular distribution of collision-related magmatism of the Hercynides, and the linear/arcuate subduction-related magmatism of the Cimmerides. Notice also that there seems to be no natural break between the two in the Balkan Peninsula and in the Caucasus. This is one reason why the Hercynides and the Cimmerides have been confused for so long in these regions. The list of references, from which this map was compiled, may be obtained from the Geological Society of London for the price of photocopying.

direct influence on the future Tethyside evolution.

The earliest 'Tethyside' (i.e. post-Devonian within the Tethyside architecture) sutures are also those between individual 'exotic' blocks of the Cimmeride collage. The earliest of these was probably the Song Ma (Red River) suture (*XXIX*) in northern Vietnam, that closed between the already accreted (?arc) fragment x' and the new (?arc) fragment x'' along a south-west-dipping subduction zone during the Tournaisian–Viséan interval (Fontaine and Workman 1978). Morgunov (1970) indicated a somewhat later time for the onset of intense south-west-directed thrusting, namely during the Viséan to Namurian transition. Recently, both Gatinsky (1985) and Hutchison (in press) argued that the closure along not only the Song Ma suture, but also the final collision between Annamia (x) and the 'South China block' *sensu lato* (w or y) had already taken place during the Devonian. The basis for this interpretation is the observation that an *c.* 10-km thick section south of the suture, consisting of Cambrian through Silurian shallow-water sediments, was deformed and uplifted before the beginning of the Devonian. A regional unconformity is believed to exist at the base of the Devonian (Fontaine and Workman 1978). However, the sea returned quickly, and the Devonian through early Carboniferous marine sedimentation was finally terminated by intense south-west-vergent folding and large-scale thrusting, during the early to middle Carboniferous, forming such major structures as the Nam Nhuong nappe (with a 220 km along-strike length and some 60 km of minimum displacement; Fromaget 1927). The presence of late Palaeozoic ophiolites along the Song Ma suture, and granites to the south of the suture (Fig. 4) further strengthens the interpretation of an early to middle Carboniferous collision along it. Farther to the north-west, along the Ailao-Shan segment of the same suture (just under the letter *v* in Fig. 3) the collision may have happened later, during the Triassic, as evidenced by a complex terrane of sedimentary and volcanic rocks plus some ophiolite slivers with ages ranging Devonian–Triassic (Bally *et al.* 1980; Şengör and Hsü 1984).

Another Carboniferous suture between two 'exotic' Cimmeride collage components is the Suolun–Xiamulun (*XVI*) suture between the North China block (*q*), and what is known as the 'North China Foldbelt' (*r*—Klimetz 1983). Yang *et al.* (1986) ascribe a middle Carboniferous age to this suture.

Finally, the Qimandag (*XII*) suture united the 'Central Meganticlinorium of the Western Kuen-Lun' block of Belyaevsky (1976) (*n*) with the Qaidam block (*o*), sometime during the late Palaeozoic, as judged from the abundant arc-type intrusive rocks and ophiolites of 'Hercynian' age along it (Fig. 4).

Whether the formation of the Qimandag suture post-dated the sutures connecting the Central Meganticlinorium of the Western Kuen-Lun (*n*) and the Qaidam (*o*) blocks with Serindia[2] or Tarim (the 'Northern Synclinorium of the Western Kuen-Lun' of Belyaevsky 1976 (*XI*) and the Arjin suture of Yang *et al.* 1986, (*XIII*) respectively), is difficult to tell so far. All are unconformably covered by Permian and younger terrestrial molasse sequences, including local Lower Triassic limestones and Middle Triassic coal (Şengör and Hsü 1984). Şengör (1984) and Şengör and Hsü (1984) interpreted the late Palaeozoic tectonics of the Western Kuen-Lun in terms of the accretion of a Central Meganticlinorium block (*n*) to the south-east margin of Tarim along suture *XI*, because of the ophiolites and granites of this age along it (Fig. 4). Recent field investigations, as summarized by Chang *et al.* (in press), have been interpreted by these authors to indicate the presence of a geometry and evolution in the Western Kuen-Lun similar to the situation encountered along the Tergun Daba Shan–Qinghai Nan Shan suture (*XVIII*), and allegedly different from the Northern Pamirs. Chang *et al.* (in press) point out that much of the northern part of the Western Kuen-Lun consists mainly of a thick sequence of marine clastic rocks, and metamorphosed intermediate and mafic submarine volcanics of Sinian–Cambrian age. This assemblage is found in association with plagiogranite-bearing ophiolites. A K–Ar age of 517 Ma was obtained from a hornblende in the plagiogranite. The metavolcanic rocks contain jadeite and quartz, indicating blueschist-facies conditions, and have led Chang *et al.* (in press) to interpret this entire rock suite as a former subduction complex. All of these rocks are unconformably overlain by fossil-bearing Ordovician rocks, which in turn are covered by a widespread Devonian molasse. This pre-Ordovician subduction complex is largely equivalent to what Norin (1946) calls 'the crystalline rocks' of the extreme Western Kuen-Lun (between 77°E, and 82°E longitudes), and what Belyaevsky (1976) designates as part of the 'Central Meganticlinorium'.

[2] The term 'Serindia' was used by Argand (1924) to denote the continental fragment forming the basement of the Tarim Basin. Herrmann (1923), however, suggested that *Serindia* was a misspelt form of *Serinda* which might be synonymous with the Arabic *Serendib*, and presented arguments to show that it was probably Ceylon.

Using the same historical data as Herrmann, however, von Richthofen (1877, pp. 528–9, 550–1) came to the conclusion that Serinda must be identical with Khotan. Since the term Serindia had a priority in the geological literature, and has been repeatedly used, by von Richthofen (as Serinda), and Stille, among others, we adopted the term (as did Şengör (1987) in an earlier draft of the manuscript), and the illustrations have been drafted with this geographical designation.

Huang (1945), called, however, the block in question Tarim. He was of the opinion that the name Tarim is better known among the Chinese. He believed that the term Serindia was an abbreviation of the expression 'Silk Roads to India', which used to run north and south of the Taklamakan Desert. The literal Chinese translation of Serindia would be *Shi Indu*, carrying the meaning Silky India. Such a translation would be incomprehensible to Chinese readers, and it would also carry the wrong implication that the area was once a part of India. Finally, Serindia as originally suggested by Argand (1924) implies that the area is underlain by a craton. Yet this is no longer a certainty, as we used to believe, now that one of us has suggested that the Tarim Desert is underlain in part by oceanic crust of a relict Palaeo-Tethyan back-arc basin (Hsü in press). For all those reasons, we now have changed all references to Serindia in the text into Tarim, except that we have left the figures as they were originally drafted.

However, the early Palaeozoic metasediments, metavolcanics, and ophiolites reported from the Western Kuen-Lun are not dissimilar to correlative assemblages known from the Northern Pamirs and the Western Hindu Kush (Boulin and Bouyx 1977, Kravchenko 1979). Farther westwards, from the Paropamisus, Wolfart and Wittekindt (1980) report similar early Palaeozoic rocks.

Thus, from the Paropamisus to the Western Kuen-Lun, and then all the way to the Tergun Daba Shan–Qilian Shan system (i.e. in regions just north of the suture *VIII*, and along the sutures *XI*, *XIII*, and *XVIII* themselves) we see clear evidence of an early Palaeozoic orogenic event, involving possibly subduction-related volcanism, ophiolite obduction and, locally, blueschist metamorphism. Whether it involved continental collision in Afghanistan, in the Northern Pamirs, and along the sutures *XI* and *XIII* is unclear, but in the east it seems to have involved the collision of a Qaidam 'block' (*o*: a collection of arcs?) with the Alxa (Ala Shan) block (*p*). Traces of this early Palaeozoic event are not encountered south of the Qaidam block in the Burhan Budai Shan and in the Anyemaqen Shan (Chang *et al.* 1986, Chang *et al.* in press).

Chang *et al.* (in press) interpret the early Palaeozoic record along suture *XI* as indicating the closure of a marginal basin, and the late Palaeozoic record as the growth of a south-facing continental margin arc/subduction–accretion complex. Although they seem unable to identify the accreted early Palaeozoic arc fragment in the Central Kuen-Lun, late Proterozoic rocks shown on the Geological Map of China (1976), just to the south-west of the Qimandag suture (*XII*) may in fact represent the basement of such a fragment.

The extant sparse data could be interpreted in one of two ways: (1) either the fragment *n* is viewed as a Palaeozoic accretionary complex, consisting essentially of a flysch/*mélange* wedge, with local, Devonian–Permian shallow-water forearc sediments, all of which function as the 'basement' to a late Palaeozoic continental-margin magmatic arc along suture *XIX*, or (2) the fragment *n* is viewed as a mini-continent—carrying a record of early Palaeozoic subduction—which collided with Tarim in late Palaeozoic times and then slid left-laterally along suture *XI* to close the Qimandag suture (*XII*) somewhat later. In this paper we adhere to the second possibility, mainly because of the presence of large masses of late Proterozoic rocks south of Qimandag and also because a well-defined belt of late Palaeozoic arc-type magmatic rocks and ophiolites everywhere separates the fragments *n* and *o* from Tarim. This belt implies active subduction, and therefore the presence of an ocean between these objects (Fig. 4) at least until the middle Carboniferous. One of us (K. J. Hsü) thinks that this subduction activity opened a major back-arc basin behind the northern Kuen-Lun-Altyn Dagh Range during the Palaeozoic. This basin is now preserved in the Tarim area, which is thought to be underlain by Palaeozoic oceanic crust (Hsü in press). This is an interesting possibility, but whether Tarim is a microcontinent or a relict back-arc basin, the Palaeo-Tethyan suture geometry and evolution around it, as presented here, remains unaffected.

The following picture thus emerges concerning the ages of the sutures in the Kuen-Lun Mountains as a whole: by middle Palaeozoic time, the Qaidam block (*o*) collided with the Alxa Massif (*p*). Later, during the late Palaeozoic, the combined Qaidam/Alxa block ($o + p$), and the block *n* independently collided with Tarim along the sutures *XI* and *XIII*, and with the Altaid Asia along the suture *XIV*, as clearly indicated by the abrupt cessation of subduction-related magmatic activity along all of these sutures after the Palaeozoic (Fig. 5). These collisions were followed shortly by the closure of the Qimandag remnant ocean.

The last sutures we discuss in the category of Cimmeride Palaeozoic sutures is the Helan Shan suture (*XXXIX*), and its apparent prolongation northward along sutures *XV* and *XVIII*. Klimetz (1982, 1983) and, following him, Şengör (1984) and Şengör and Hsü (1984), considered the Helan Shan as a middle Mesozoic suture zone, mainly because of an erroneous identification of blueschists in the Jurassic Siasouchan Series (Lexique Stratigraphique International 1963). The main—and a very strong—argument against the presence of a middle Mesozoic suture along the Helan Shan, is the absence of any Mesozoic marine sediments. It may possibly represent a late Palaeozoic suture (or late Proterozoic according to some, e.g. Sobolev 1984) that was later reactivated to accommodate further intracontinental shortening—probably a substantial amount, as can be judged from the intensity of folding in the Jurassic sediments (Teilhard de Chardin and Licent 1924).

A problem similar to the one encountered in the Helan Shan case, also exists along the only major Palaeozoic marginal suture belt separating Altaid Asia from the larger fragments of the Cimmeride collage, namely the Inner Mongolian–Da Hingan (Greater Khingan) suture (*XV*, *XVII*) (Wang and Liu 1986), which

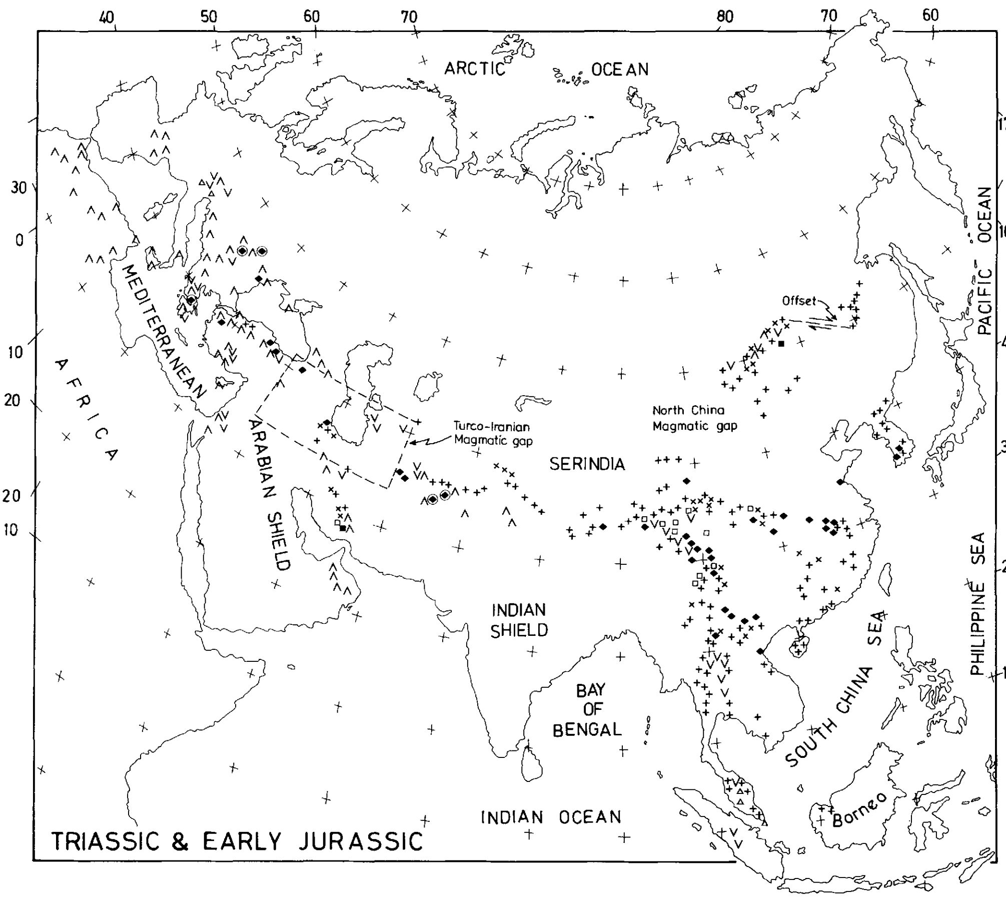


FIG. 5. Triassic and early Jurassic ('Indosinian' in China) magmatism in the Tethysides. The list of references, from which this map was compiled, may be obtained from the Geological Society of London for the price of photocopying. For legend see Fig. 4.

continues directly into the Shilka suture (*XLI*). A well-developed early Cretaceous magmatic arc, and continental-margin sediments along the Shilka suture (Kosygin and Parfenov 1981), leave no room for doubt for the existence of a late Mesozoic suture here. Farther south, however, despite some Triassic and early Jurassic, and very intensive and widespread late Jurassic to early Cretaceous intermediate and felsic magmatism (Figs 6 and 7), neither coeval marine sediments nor any ophiolites, are seen. Although Şengör (1984) interpreted the ultramafics located generally to the east of the extensive Mesozoic felsic magmatic rocks, as indicating the existence of a coeval suture zone, this interpretation can no longer be upheld, because the associated ultramafics all seem to be late Palaeozoic in age.

Convergent plate-margin magmatism appears to be confined to the Greater Khingan–Inner Mongolian suture zone during the late Palaeozoic (Fig. 4). This suture probably closed at this time, as the late Palaeozoic ophiolites and the nearly complete cessation of all magmatism bear witness (Fig. 5) (see also Wang and Liu 1986). Immediately following this closure, the locus of convergence seems to have jumped to the Mongolo-Okhotsk ocean and to have begun consuming it. Following the final elimination of the Mongolo-Okhotsk ocean in the late Jurassic, continuing intracontinental convergence generated the wide felsic magmatic area, and probably also triggered subduction along the Ussuri suture in the Sikhote Alin (Fig. 6). In this episode, the Greater Khingan suture appears to have been reactivated in a compressional mode, and this reactivation extended its effects probably as far south as the Helan Shan (Wong 1929).

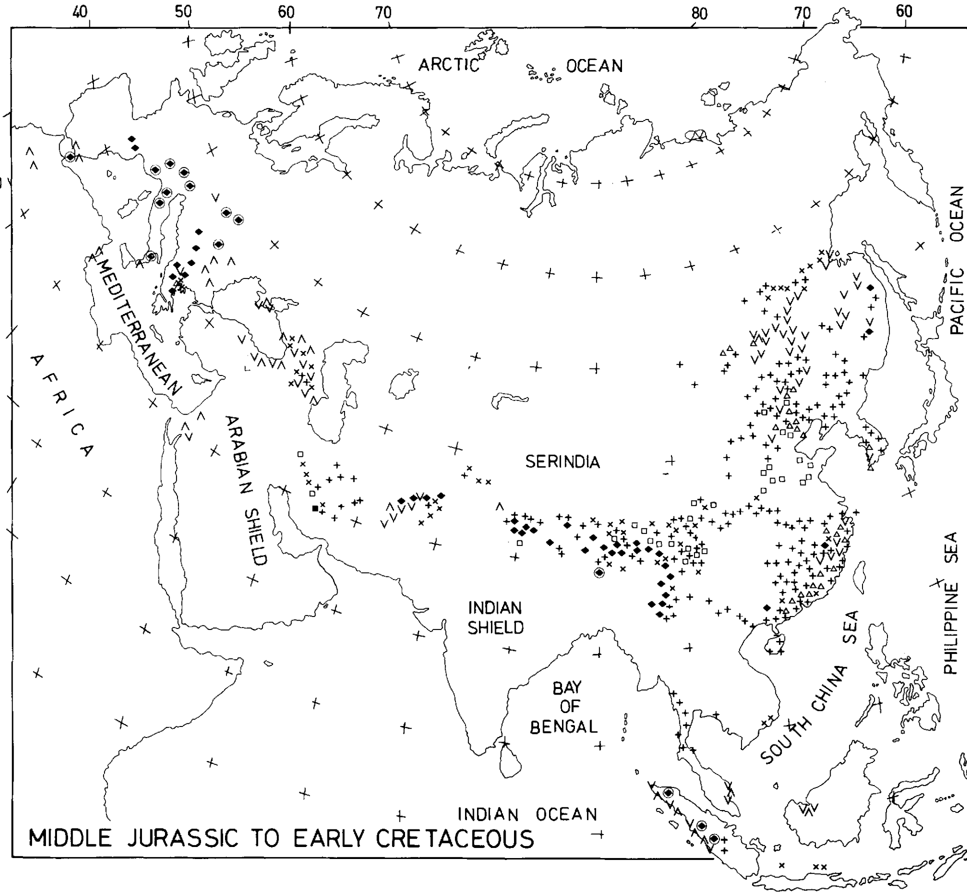


FIG. 6. Middle Jurassic to early Cretaceous ('early Yenshenian' in China) magmatism in the Tethysides. The list of references, from which this map was compiled, may be obtained from the Geological Society of London for the price of photocopying. For legend see Fig. 4.

Mesozoic Cimmeride sutures

As the 'North China magmatic gap' in Fig. 5 also shows, all oceanic plate boundaries that connect north-east China with the rest of the Tethyside system, had become extinct by the beginning of the Mesozoic, as a consequence of late Palaeozoic continental collisions in this part of the world (Wang and Liu 1986). The Mesozoic Pacific-type, active continental margin thus became more uniformly south-facing (present geographical orientation), along which magmatic arc activity continued without break, and very much in the same places as in the late Palaeozoic (compare Figs 4 and 5). The only significant exception was in the Central Kuen-Lun, where the magmatic arc axis jumped a few hundred kilometres southwards, presumably after the accretion of the Central Meganticlinorium (*n*) to Tarim.

The earliest Mesozoic continental collisions again appear to have been between individual components of the Cimmeride collage. The first of these was probably the collision along the Jinsha Jiang suture (*XXI*), between the eastern segment of the North Tibetan Qangtang block (*s″*) and the Chola Shan sliver (*v*). Along the Hoh Xil Shan Arc (to the west of the Jinsha Jiang suture) magmatic activity lasted until the Carnian, and then the suture zone was uplifted. The marine Triassic section along it commonly ends with the Carnian (Şengör and Hsü 1984).

A probable coeval suture is the Song Da (Black River) suture (*XXX*) in North Vietnam. Along the early Carboniferous Song Ma suture, which closely parallels the later Song Da suture,

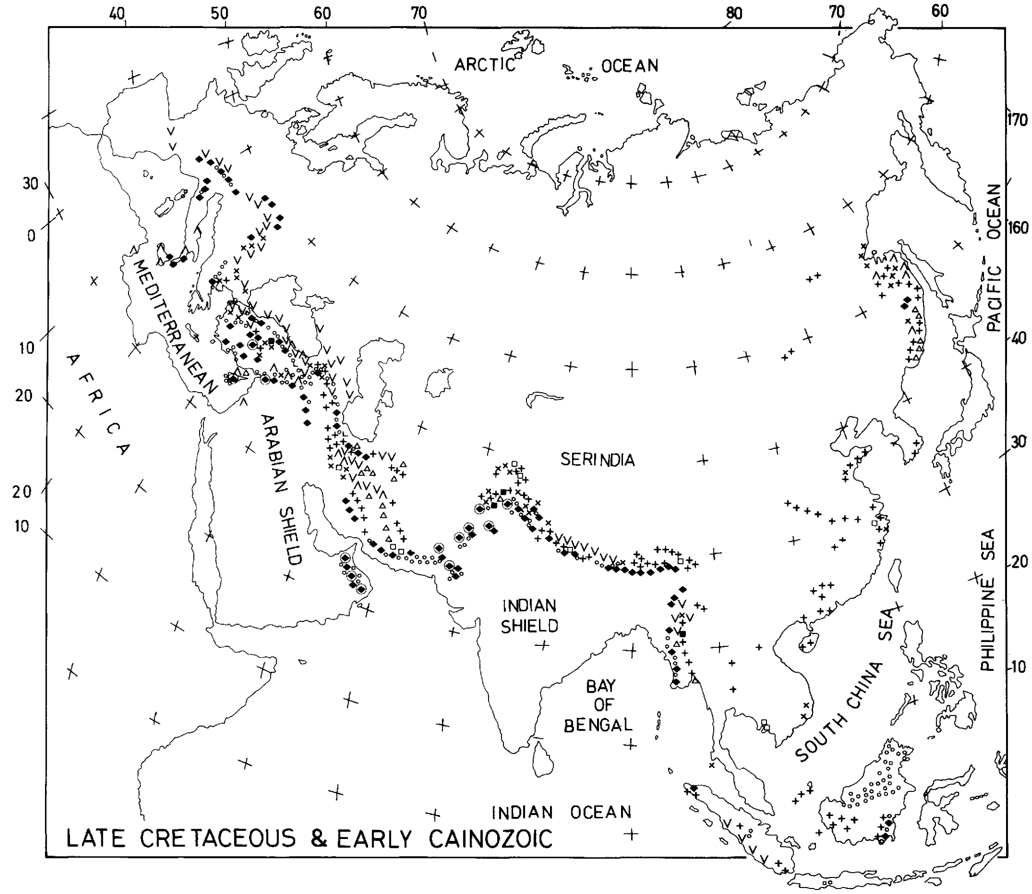


FIG. 7. Late Cretaceous and early Cainozoic ('late Yenshenian' and partly 'Himalayan' in China) magmatism in the Tethysides. References, from which this map was compiled, may be obtained from the Geological Society of London for the price of photocopying. For legend see Fig. 4.

the post-deformation rocks are almost exclusively middle Carboniferous to Permian shallow-water limestones, similar to the Devonian rocks deposited following the arc–continent collision along the Tamky–Phueson suture (*XXVIII*) to its south. This facies suggests that the early Carboniferous deformation along the Song Ma suture was likely due to the collision with Annamia of yet another island arc, now perhaps buried under the younger deposits of the Cainozoic Song Ma rift zones. That the Song Ma collision did not represent the terminal welding of Annamia to South China, is suggested by the supposedly 10 000-metre thick, pre-Norian Triassic section, which contains abundant volcanic material such as spilites, albitophyres, basalts, plus *ultramafics* (Salun *et al.* 1975) along the Song Da suture. These rocks are now extremely deformed and imbricated, with a clear north-east vergence under the unconformable coal-bearing Norian–Rhaetian molasse (Fromaget 1934; Burrett 1974). Şengör (1984) interpreted them as a subduction–accretion complex, in which measured stratigraphic thicknesses would have little meaning.

To the north-east of the Song Da suture, on its foreland, the Hagian–Bac Kan nappe carried Precambrian metamorphic basement rocks on to folded late Permian limestones, during the late Triassic. This episode ended with the 'intrusion' of the Viet Bac mantled gneiss dome into the Hagian–Bac Kan nappe (Fontaine and Workman 1978; Şengör and Hsü 1984). Triassic and Jurassic convergent-margin magmatism, in the form of dominantly granitic intrusions (Figs 5 and 6) also indicates that the final collision along the Song Da suture could not have been much earlier than the late Carnian.

Additional support for this view comes from the Ailo Shan, in the Chinese province of Yun-

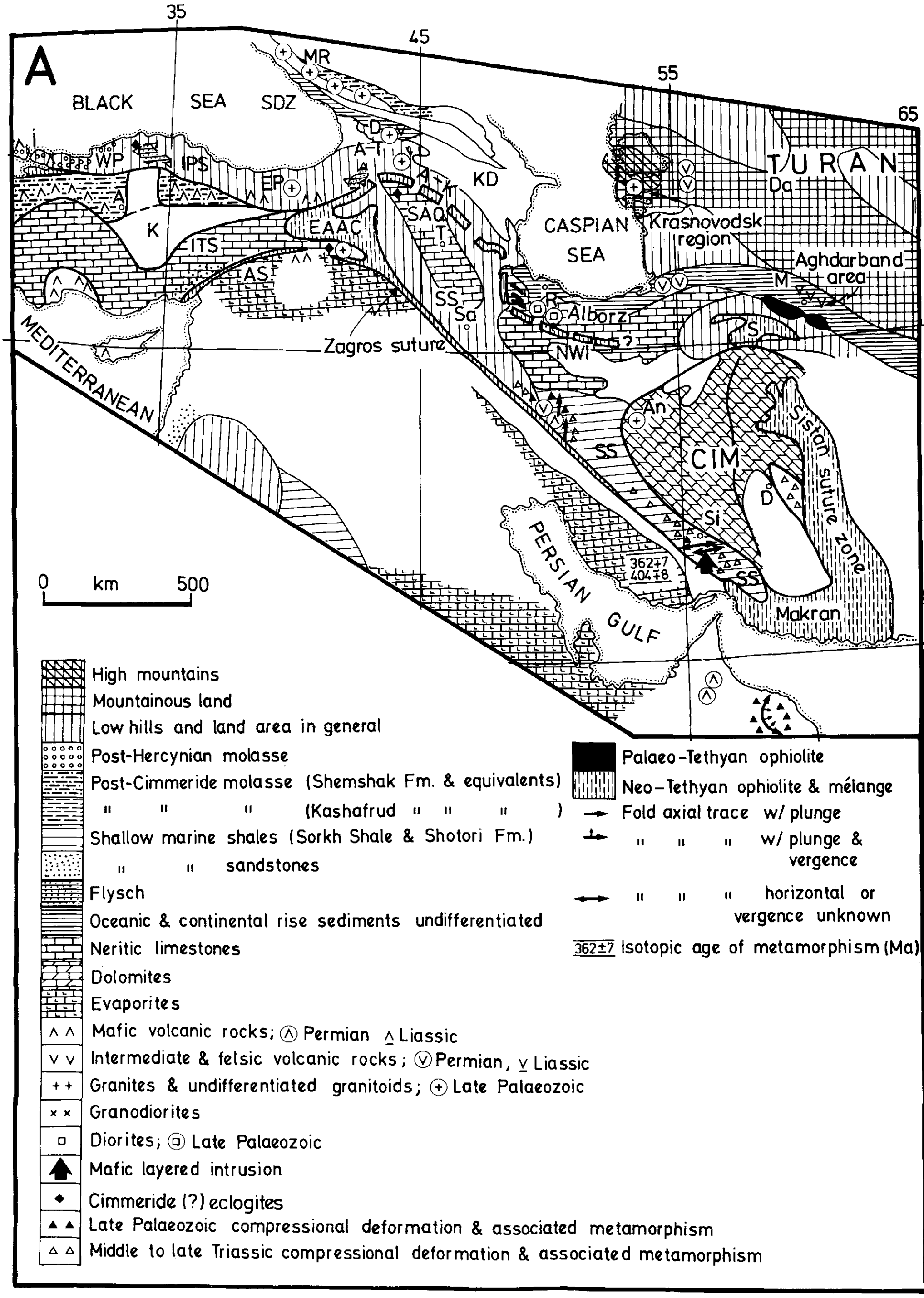


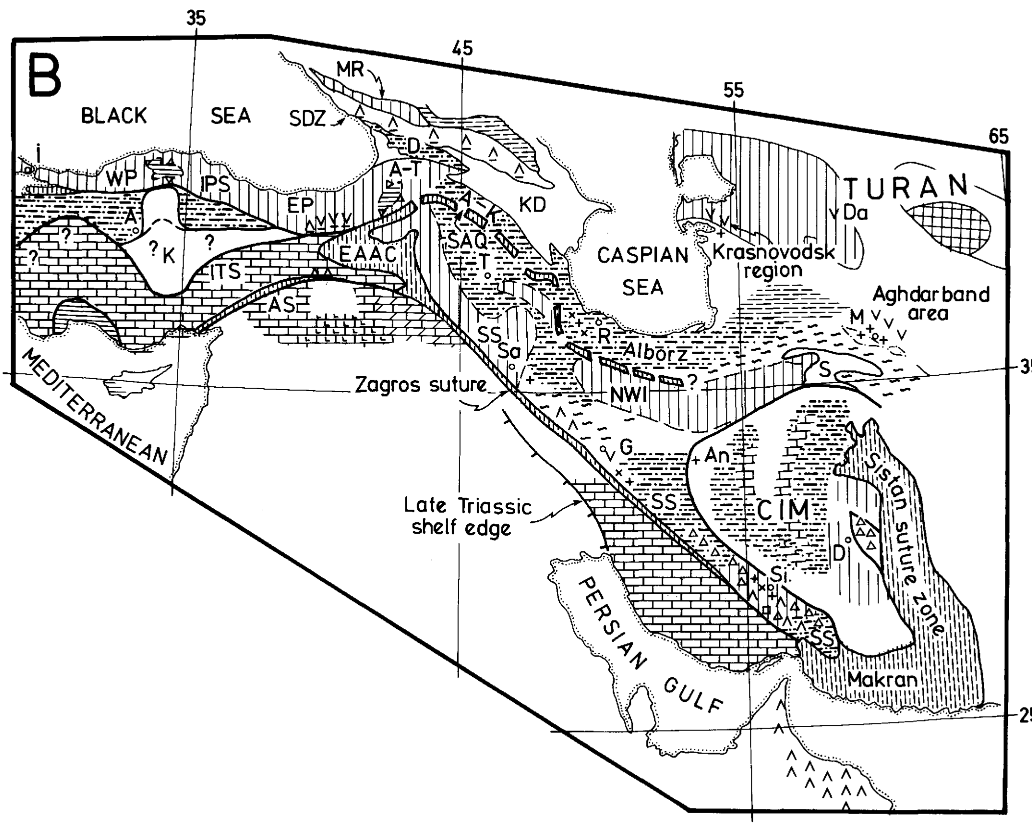
FIG. 8. (A) Distribution of major early and middle Triassic sedimentary facies, igneous rocks, metamorphism, and deformation in Iran and surrounding regions. Compiled from Thiele *et al.* (1968), Vereshchagin and Ronov (1968), Lunnov (1971), Vialon *et al.* (1972), Khain (1975), Stämpfli (1978), Berberian and King (1981), Şengör and Yılmaz (1981), Michard (1983), Roman'ko and Morozov

nan (see suture segment *XXII*). In these mountains there is a 10- to 15-km wide ophiolitic *mélange* zone of middle Triassic age. Late Triassic red beds and felsic volcanics unconformably overlie the *mélange* (Bally *et al.* 1980).

The initial collisional contact established between Laurasia and the Cimmerian Continent occurred in Iran, and represents the next oldest Cimmeride suturing event. Following Stöcklin (1974), both Hsü (1977) and Şengör (1979*a*, 1984, 1985*a*, *b*; 1986*a*) argued that the ophiolite slivers encountered under the unconformable Rhaeto-Liassic terrestrial Shemshak Formation, in the Alborz/Binalud ranges in northern Iran, near the towns of Rasht (Davies *et al.* 1972; Clark *et al.* 1975) and Mashhad (Alavi 1979), were remnants of Palaeo-Tethys. They further argued that the Palaeo-Tethys in Iran had closed before the Rhaetian and after the Carnian, because the platform deposits, on which the Mashhad ophiolites had moved, reached as far upwards as the Carnian (Alavi 1979).

None of these authors, however, could account for the strong, coeval compressional deformation, kyanite-grade metamorphism and arc-related volcanic and intrusive activity of Devonian to late Triassic age along the *south-eastern* border of central Iran, in the Sanandaj–Sirjan zone (*i*) under unconformable Liassic rocks of the Shemshak Formation equivalents (*SS* in Figs 8a and b).

Berberian and King (1981) recognized the necessity of assuming convergent plate-margin tectonics along the Sanandaj–Sirjan zone in the late Palaeozoic Triassic interval, and postulated the subduction *northward* under the Sanandaj–Sirjan zone of a Palaeozoic 'Zagros ocean'. However, they also admit that stratigraphic, palaeobiogeographical and palaeomagnetic data unequivocally demand that Iran be kept as an



(B) Distribution of major late Triassic and earliest Jurassic sedimentary facies, igneous rocks, metamorphism, and deformation in Iran and surrounding regions. Sources as in Fig. 8a. Symbols are as used in Fig. 8a.

(1983), Adamia and Belov (1984), Davoudzadeh and Schmidt (1984), Meliksetyan *et al.* (1984), Ruttner (1984), Şengör *et al.* (1984*b*), Okay *et al.* (1985).

integral part of Gondwana Land until the early Triassic, so that no ocean could have existed between it and Arabia, prior to at least the middle Triassic. Moreover, Michard (1982, 1983) described folded and slightly metamorphosed terrains, constituting the pre-Permian basement of the Oman Mountains, and ascribed their origin to a late Palaeozoic orogeny (Fig. 8a). Thus, deformation and metamorphism similar to those known from the pre-Liassic basement of the Sanandaj–Sirjan zone, also took place on the 'Arabian' side of Berberian and King's (1981) Palaeozoic Zagros ocean, which cannot be explained by northward subduction under the Sanandaj–Sirjan zone.

There is also abundant evidence that the Zagros segment of the Neo-Tethys, between Iran and Arabia, began rifting during the late Permian (e.g. Szabo and Kheradpir 1978), with continuing stretching and abundant basaltic volcanism in the early and middle Triassic (Searle *et al.* 1980, Searle and Graham 1982), and finally the onset of ocean-floor spreading in the late Triassic.

The late Palaeozoic and Triassic (locally later) tectonism in the Sanandaj–Sirjan zone is very reminiscent of the events known from the Aghdarband/Kopet Dagh regions of north-east Iran and south-west USSR (Turkmenia SSR; along suture *VII*). In the Krasnovodsk uplift of the Kopet Dagh trend, there was both late Palaeozoic (mainly Permian), and Triassic arc-type magmatism, whereas in the Aghdarband uplift, this activity is seen only in the Triassic (Figs 4, 5, and 8a). On the basis of borehole and geophysical data, Bazhenov and Burtman (1986) traced a continuous late Palaeozoic magmatic arc, composed mainly of andesites and dacites, between Krasnovodsk and Aghdarband. Stämpfli (1978) reported Permian volcanics from the Gonbad-ı Qabus area, near the south-east corner of the Caspian Sea (Figs 4 and 8a). Between the Sanandaj–Sirjan zone and the Alborz Mountains, the entire Iranian area contains no comparable deformations or magmatism, except in the Anarek area (Davoudzadeh *et al.* 1981, Roman'ko and Morozov 1983) (Fig. 8a and 8b).

Within the framework of the present-day geography it is thus difficult to find a suitable convergent plate boundary of Permo-Triassic (and older) age, to create the Sanandaj–Sirjan magmatic arc belt. An examination of Figs 4 and 5 shows the continuous late Palaeozoic–early Triassic magmatic arc belt—related to the consumption of Palaeo-Tethys in Central Asia—to stop rather abruptly in the Aghdarband area. This interruption, the Turco–Iranian magmatic gap (Fig. 4), is partly related to younger cover in the Turkmenian SSR, hiding the segment of the arc between Aghdarband and Krasnovodsk (see Bazhenov and Burtman 1986; Fig. 2). But farther west from the Krasnovodsk, in the Greater Caucasus, we see no trace of the Triassic arc. Instead, a long but seemingly isolated arc segment is present in the Sanandaj–Sirjan zone, that continues into the Pontides in northern Turkey (Figs 5, and 8a and b). The apparent interruption in eastern Turkey, visible in Figs 4 and 5, is due to the presence of a late Cainozoic volcanic cover (Şengör and Kidd 1979).

This haphazard geometry, involving abrupt termination of major fossil magmatic arc systems (as in Krasnovodsk), presence of fragments of others, apparently with no associated sutures (as in the Sanandaj–Sirjan zone), and very small pieces of former arcs lying in complete isolation amidst anomalous palaeogeographical zones (as in the Anarek region of Central Iran; Figs 4, 5, 8a, and 8b) suggests the disruption and redistribution, by strike-slip faulting, of a formerly more 'conventionally' organized orogenic belt. This view is strengthened by the 135° anticlockwise rotation of the Central Iranian Microcontinent (*k*) between the late Triassic and the Cretaceous (Davoudzadeh *et al.* 1981; Soffel and Förster 1984). As a result of this rotation, the originally south-east end-segment of the Sanandaj–Sirjan zone was torn away and transferred to eastern Iran, to the east of Deh Salm (Davoudzadeh and Schmidt 1984), whereas the Anarek region, originally close to the northern Palaeo-Tethyan suture near Mashhad, was brought westward to its present location (Davoudzadeh *et al.* 1981) (Figs 8a and 8b). This is also documented by the rotation of Cambrian (Zarkhov 1984), Devonian (Weddige 1984), and Mesozoic sedimentary facies (Davoudzadeh and Schmidt 1984) within the Central Iranian Microcontinent (see Fig. 8a and b for the rotation of the middle Triassic to earliest Jurassic facies realms).

Geological information from Iran alone is insufficient to reconstruct with confidence the pre-disruption geometry of the Iranian portion of the Cimmeride orogenic collage. We therefore defer the presentation of our reconstruction of how Iran evolved during the Mesozoic, until the end of our discussion on the distribution of Tethyside sutures and blocks.

A probably coeval collision occurred sometime between the middle Carnian and the Rhaetian in Thailand, along the Nan-Uttaradit–Sra-Kaeo suture (*XXVII*) (Bunopas 1985; Şengör 1986*b*), which continues southwards into the Malay Peninsula where it becomes the Bentong–Raub suture (*XXXI*) which closed after the Ladinian and before the deposition of the late Triassic Tembeling Formation (Şengör 1984, 1986*b*). Farther south, this suture crosses the Makassar Strait and separates northern Sumatra from southern Sumatra (*XXXI'*) (Hamilton 1979, Hutchison in press).

Another late Triassic suture belt within the Cimmeride collage is the Qin-Ling (*XXIII*)/ Dabie Shan (*XXXIII*)/northernmost Jiangsu (*XXXIII'*)/Chugareong (*XXXV*)/system, that united the North China platform (*q*) with the Yangtze block (*w*). Of this long suture belt, the Qin-Ling/Dabie Shan segment (*XXXIII*) is the best-known part.

Controversy however persists, both on the timing of suturing (e.g. Şengör 1985*c*) and on the predominant vergence (e.g. compare Xu *et al.* 1984, with Zhu *et al.* 1983 and Liu and Li 1984) of this complex and much disrupted orogenic belt. During the last two decades, opinions on the age of the main orogeny along the Qin-Ling have varied from Devonian (*c.*400 Ma) (e.g. Mattauer *et al.* 1985) to late Triassic (*c.*220 Ma) (e.g. Şengör and Hsü 1984; Şengör 1985*c*) or even early Jurassic (*c.*200 Ma) (Opdyke *et al.* 1986). Şengör (1985*c*) summarized the observations supporting a late Triassic collision.

Further evidence for a Triassic collision is also found farther east along the strike, in the Dabie Shan. Fig. 9 shows a cross-section across the Dabie Shan near its eastern end, where the entire orogenic belt is cut and offset some 500 km northward by the Tan-Lu Fault (Fig. 10) (Xu and Lu in press). This cross-section shows that most of the Dabie Shan consists of large slabs of Precambrian basement rocks (Dabie Group, with an age ranging from early Proterozoic to late Proterozoic, and with scattered K–Ar cooling ages as young as 532 Ma: Xu *et al.* 1984*a* and Xu Shutong pers. comm. 1986) separated by *mélange* zones, where ultramafic rocks, mafic volcanics, and marble blocks are seen scattered in a metapelitic matrix. The metamorphism of these *mélange* zones ranges from greenschist to high amphibolite grade, with local anatectic melting. Both within the Dabie Group and in the *mélange* belts, eclogites of different ages have been encountered (see Xu *et al.* 1984*a* for a map). Recently, Li *et al.* in press reported Sm–Nd isotopic ages from the eclogites and from the ultramafic rocks found in the Northern Ultramafic Belt of the Dabie Shan (Fig. 9). These ages are 243.9 ± 5.6 Ma for the eclogite, and 230.6 ± 30.7 Ma and 402.6 ± 17.4 Ma for the ultramafics, suggesting that the ultramafics formed at two different intervals (Triassic and early Devonian), and that subduction was under way already during the earliest Triassic. Moreover, Sang *et al.* (1986) have dated the metavolcanics of the Susong Group in S. Dabie Shan (Figs 9 and 11) and have concluded that the volcanics were formed during the Ordovician, and that following a later episode of metamorphism they were cooling by 231 Ma ago. Sang *et al.* also obtained younger cooling ages around 210 Ma. Stratigraphic indication of a post-Palaeozoic continental collision in the Dabie Shan, comes from the presence of a fairly thick Palaeozoic continental margin sequence: the Fozhilin Group, which was strongly deformed, with a northerly vergence, after the ?Permian and before the Jurassic (Zhu *et al.* 1983).

Fig. 11 shows our interpretation of the evolution of the Dabie Shan segment of suture *XXXIII*. This suture probably continues through the widespread ultramafic and mafic rocks of northernmost Jiangsu (XXXIII') (Chinese Academy of Geological Sciences 1973) and the 'Chugareong zone' of Lee (1984), located south of the Permo-Triassic Pyeongan Basin (Lee 1974).

The Qin-Ling/Dabie Shan mountain range delimits a continental region to the north, known as the South China platform or the South China block in the literature (Fig. 10; e.g. Klimetz 1983, Lin *et al.* 1985*a*). Huang (1978) separated the South China block into a Yangtze paraplatform in the north and east, and a South China fold system in the south and west. A much smaller tectonic unit, called the 'south-east maritime fold system', was distinguished by Huang (1978), east of the South China fold system and separated from it by the Lishui–Haifeng 'depth fracture' of Palaeozoic age (Fig. 10). Huang views the South China fold system as a unit that formed during the early Palaeozoic ('Caledonian' as he calls it), and the south-east maritime fold system he regards as being late Palaeozoic ('Variscan') in age. In 1981, Hsü pointed out that the notion of a 'South China Platform' is erroneous and misleading, owing to extensive Mesozoic deformation of the 'platform sediments' and coeval convergent-margin magmatic activity. Hsü (1981) also indicated that South China was greatly similar to the Central and Southern Appalachians, comparing the flat-lying Mesozoic strata of the Sichuan Basin (Fig. 10) with the Alleghany Plateau, and the narrow folds involving Mesozoic strata (between SP and TY in Fig. 10) with those of the Valley and Ridge Province. Farther to the south-east of these folds, are extensive outcrops of a thick unfossiliferous sequence of slates and turbidites, forming generally mountainous terrain. Blocks of metamorphosed mafic volcanics

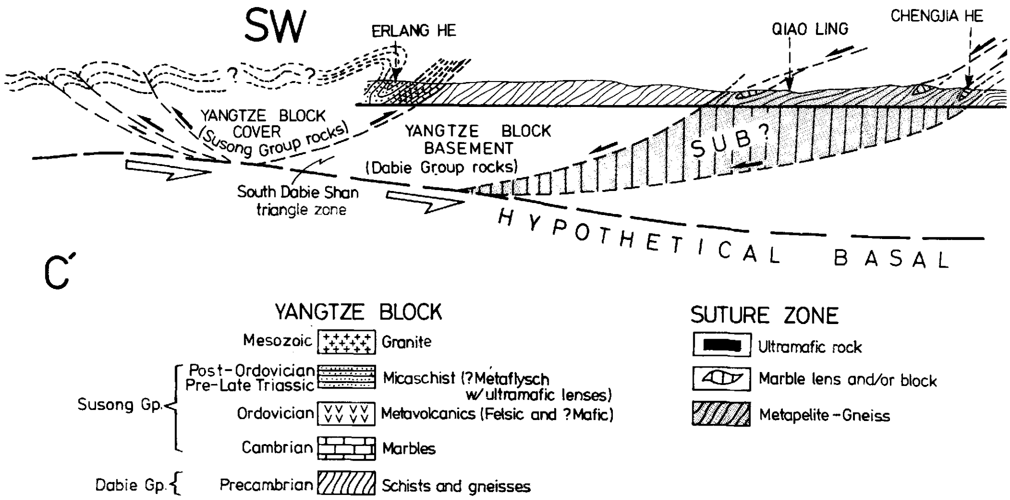


FIG. 9. An interpretive structural cross-section across the eastern end of the Dabie Shan, between Songshan Zai and Erlang He (for location see Fig. 10). Data were reinterpreted from the Taihu (H–50–XV, 1970), Lioan (H–50–III, 1974), and the Yuexi (H–50–IX, 1974) sheets of the 1:200 000

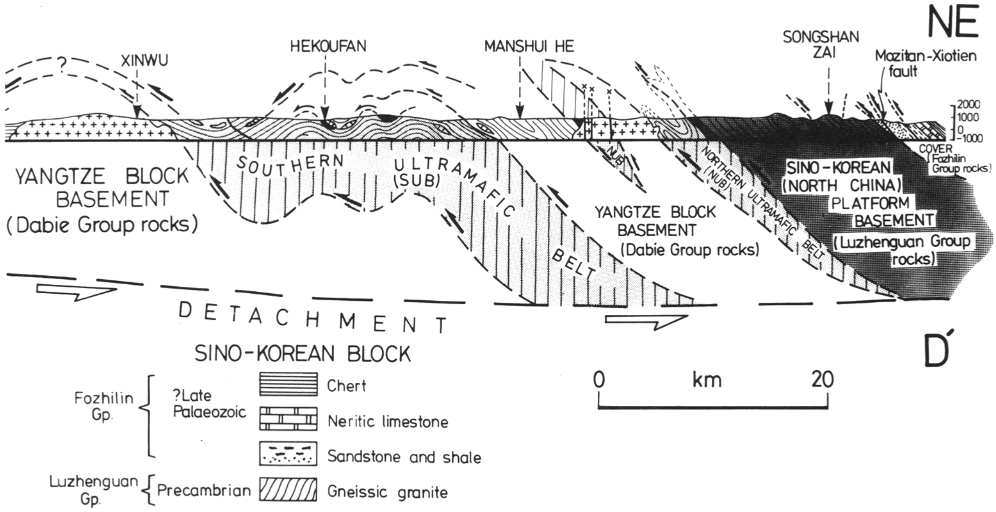
('greenstones'), submarine tuffs, gabbro, and serpentinite are locally present within the dominant slate matrix, with few chert lenses. In the provinces of Hunan and Guangxi, this complex is known as the Banxi Group (Fig. 10), and has been assigned an Upper Proterozoic age (Chinese Academy of Geological Sciences 1973). Hsü *et al.* (in press) extended this appellation to all such rocks in the South China Platform, and interpreted them as lying over the autochthonous Yangtze carbonate rocks and clastics above a thrust contact (Fig. 10). Hsü *et al.* (in press) argue that these allochthonous flysch/*mélange* masses were expelled from what they term the Xinagganzhe suture (from the abbreviated Chinese expressions for Hunan, Jiangxi, and Zhejiang), that coincides with the boundary between Huang's (1978) Yangtze paraplatform and the South China fold system tectonic units (*XXXVI* and *XXXVII*).

The Xiangganzhe suture itself actually consists of two separate segments (*XXXVI* and *XXXVII*), whose relations to each other are as yet unclear. In the north, it starts near the town of Shaoxing and extends south-westward to Jiangshan (Shui *et al.* 1986). This north-west-vergent segment—contained entirely within the Zhejiang province—continues directly into the Pingxiang–Yushan segment in Jiangxi. We will call these two segments (collectively) the Shaoxing–Pingxiang suture (*XXXVI*; *SP* in Fig. 10). To the north-west of this suture, large outcrops of the Banxi are themselves tectonically overlain by Precambrian plutonic and metamorphic basement rocks forming prominent topographic highs.

In the province of Hunan the continuity of the suture zone is interrupted; and it jumps some 500 km westwards to continue as the Tianyang suture zone (*XXXVII*; *TY* in Fig. 10). To the west of the Tianyang suture we see again an enormous mass of Banxi Group rocks thrust westward.

That the Xiangganze structure represents a major suture zone in its various segments is now generally acknowledged (e.g. Jin and Fu 1986, Shui *et al.* 1986), but the timing of suturing and the expulsion of the Banxi allochthons remains in dispute. Although the Banxi rocks have long been considered unfossiliferous, Hsü *et al.* (in press) quote unpublished information gathered by the Jiangxi Geological Survey about Permian spilites and radiolarites included in the Banxi rocks in the Jiangxi province (i.e. north-west of the Shaoxing–Pingxiang suture: SP in Fig. 10). To the west of the Tianyang segment, Yang *et al.* (1981) reported the existence of Permo-Triassic radiolarian cherts and spilitic volcanics (as in Jiangxi farther to the north-east) the former containing a 'Tethyan' fauna, including such forms as *Otoceras* and *Ophiceras*. These rocks are seen to be associated with flysch deposits that form part of the Banxi allochthon in eastern Guizhou and northern Guangxi.

The sparse biostratigraphic information thus suggests that the Banxi rocks are at least partly



geological map of the Anhui province and Xu *et al.* (1984, and in press), locally supplemented by our own observations.

as young as the early Triassic, and therefore their thrusting westwards and the suturing of the Xiangganzhe branch of Palaeo-Tethys could not be earlier than that date, at least in its south-western one-third.

In order to establish the upper limit for the age of closure of the Xiangganzhe ocean, Şengör and Hsü (1984) plotted the distribution of Palaeozoic and Mesozoic granites to the east of its suture zone. In south-east China there seem to exist *three* Mesozoic intrusive, and subordinate volcanic, belts parallel with the Xiangganzhe suture: two I-type granite belts flanking a much wider S-type granite belt. Şengör and Hsü (1984) interpreted the western I-type granite belt and a considerable portion of the S-type belt as manifestations of a south-east-dipping subduction under the Huanan block (*y* at large). On the Yangtze block (*w*) and on the Huanan block the post-orogenic molasse sediments are late Triassic or younger (Chinese Academy of Geological Sciences 1973; Hsü *et al.* in press) and we therefore place the suturing between these two blocks in the middle Triassic, with intracontinental convergence continuing until early Cretaceous times, and causing both the formation of the wide, S-type, 'Tibet-style' intrusive belt and the widespread middle Jurassic to middle Cretaceous compressional intramontane basins.

Although major suturing along the entire length of the Xiangganzhe suture zone was a middle Triassic affair, earlier deformation, and the deposition of terrestrial 'molasse' sediments are known from the Huanan block in Devonian and pre-Devonian times. A Devonian 'clastic wedge' thins to the west, suggesting that it was fed from an easterly source (Hsü *et al.* in press). Grabau was the first to notice this, and he called the hypothesized source area the 'Cathaysia' landmass. Hsü *et al.* (in press) followed Li *et al.* (1982) in interpreting the Lishui–Haifeng depth fracture of Huang (1978) as a pre-Devonian suture zone (*XXXVIII*) that united the 'southeast maritime fold system' (*y*″: 'coastal block' of Hsü *et al.* in prep.) with the Huanan block proper (*y* in the restricted sense).[3] Although Devonian clastic rocks also exist on the Yangtze block, they are not unconformable on older rocks. Here also, they do not have a typically molasse character. A possible early Palaeozoic deformation, affecting the Yangtze block, may be indicated by middle Ordovician turbidites with a greywacke composition, in the Lin'an county of the Zhejiang province to the north of the Shaoxing–Jiangshan suture zone.

Fig. 12 represents the interpretation of Hsü *et al.* (in press), of the evolution of what Hsü *et al.* (in prep.) call the 'Cathaysian orogenic belt' and we adopt it here also. The Palaeozoic granites, shown intruding the Huanan block in Fig. 12, have been partly dated recently and are located in the provinces of Hunan, Fujiang, Guangdong

[3] Recent observations by Professor J. L. Li in the Fujiang province suggest that at least parts of the Lishui–Haifeng structure may represent Mesozoic sutures.

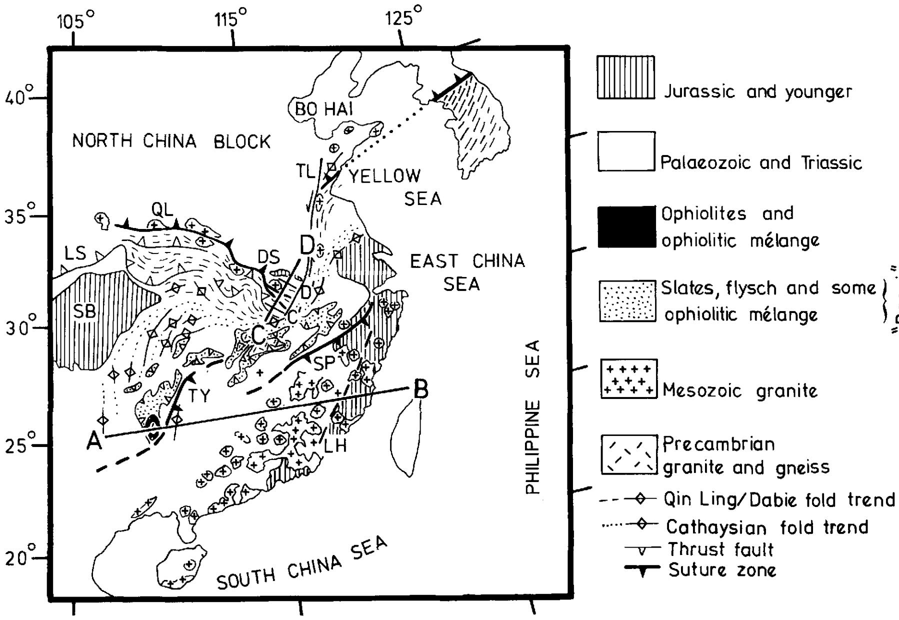


FIG. 10. Simplified tectonic map of South China and its proposed correlation with Korea (compiled from the Geological Map of China 1976, Lee 1984, and Hsü *et al.* in press). Key to lettering: *C*—Chugareong suture zone, *DS*—Dabie Shan, *LH*—Lishui–Haifeng suture zone, *LS*—Longmen Shan, *QL*—Qin-Ling, *SP*—Shaoxing–Pingxiang suture zone, TL—Tan-Lu Fault, TY—Tianyang suture zone. *C'–D'*, C–D, and A–B show the locations of the cross-sections shown in Figs 9, 11 and 12 respectively.

and eastern Guangxi (Fig. 4; Institute of Geochemistry 1980).

Before we close our discussion of the Cimmeride tectonics of South China we should mention the possibility of the presence of an independent Hunan block (y') in the architecture of South China as suggested in Fig. 3. Alternatively, the free ends of the Tianyang and the Shaoxing–Pingxiang sutures may have been originally contiguous and later offset by a post-collisional, right-lateral, strike-slip fault (Jin and Fu 1986). This latter interpretation seems also to be favoured by the presence of several north-north-east trending, narrow, fault-controlled Cretaceous depressions.

With the discussion of the Cimmeride assembly of South China, we have come to the end of our description of the Triassic suture zones within the Tethysides. However, there is one more major region within the Cimmerides where a large number of sutures are present and where marine conditions largely ended by about the end of the Triassic. This region, which has been interpreted as an 'Indosinian' zone of consolidation, because of the retreat of the sea from it during the late Triassic (e.g. Chen 1977; Huang 1978; Li 1980; Li *et al.* 1982), is the Songpan–Ganzi System (Fig. 4), the giant suture knot of the East Asian Cimmerides (Şengör 1981, 1984).

Şengör (1981, 1984, 1986*b*) and Şengör and Hsü (1984) discussed the geology of this extremely complex and little-known region. As shown in Fig. 3, the south-west margin of the Songpan–Ganzi System consists of a north-east-convex bundle of four suture zones. In reality, the outermost of these, the Litang *mélange* zone (*XXIII*), is not a suture but simply a structural boundary, marked by a few 'Indosinian' ultramafics separating the Shaluli Shan Arc system (u) from the thick Triassic flysch fill of the main mass of the Songpan–Ganzi System. This entire flysch fill, the Bayanhar Group, represents a suture knot itself (see Şengör 1981, 1984, 1985*a*, 1986*b*).

A second belt of Indosinian ultramafics (largely Triassic) representing the Maniganggo suture (*XXII*), separate the Shaluli Shan Arc from the Hoh Xil Shan Arc constructed directly on top of the basement of the western part of the western segment of the Qangtang block (s'') between 88°E and about 97°E; farther east a new zone of dominantly Jurassic–early Cretaceous granitic and granodioritic plutonic activity begins inserting itself between the Shaluli Shan Arc and the Hoh Xil Shan continental-margin arc. This new zone is termed the Chola Shan Arc (v) by Şengör (1984). The Maniganggo suture also separates the Chola Shan from the Shaluli Shan. Between the Hoh Xil Shan Arc and the

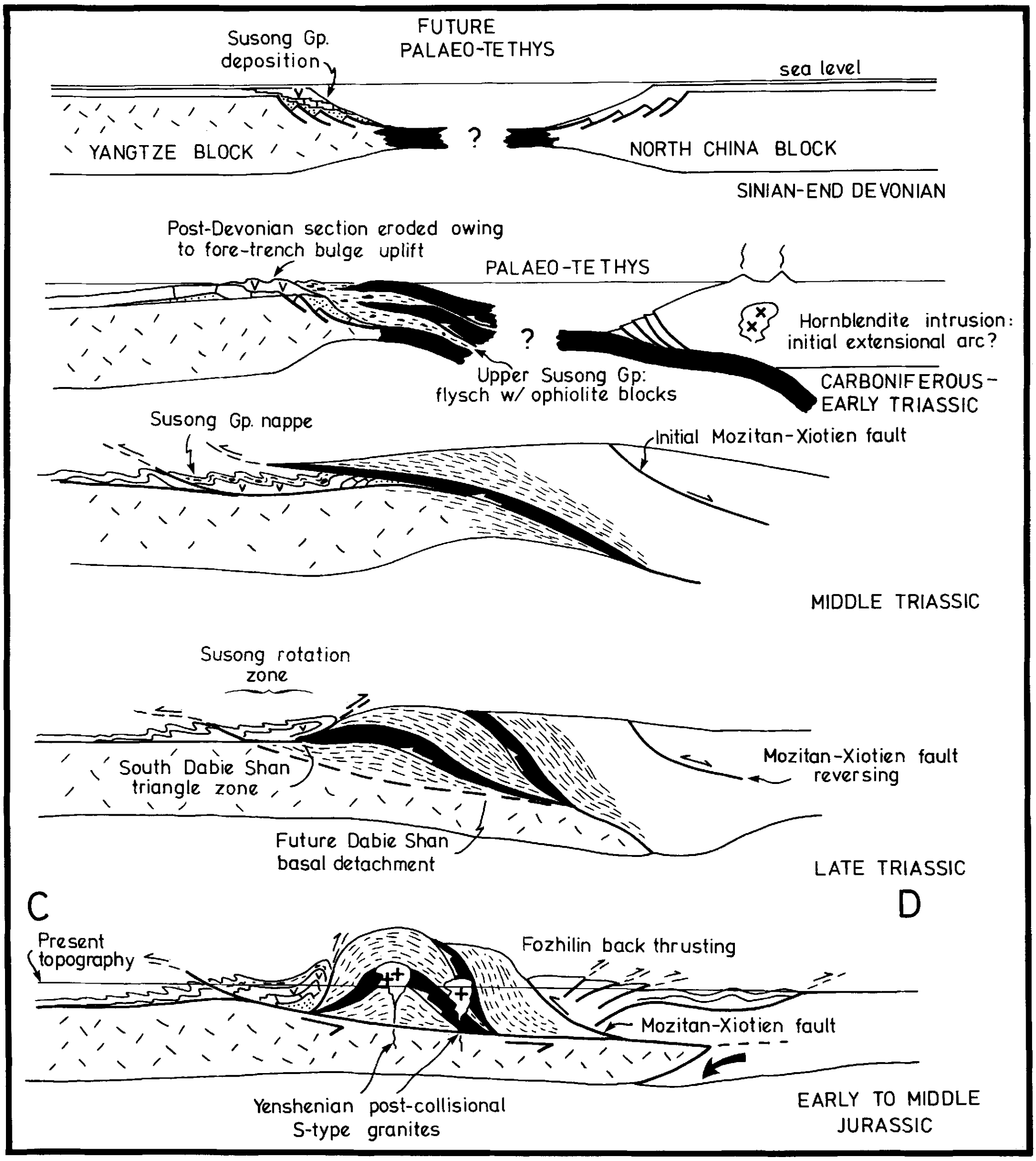


FIG. 11. Schematic, sequential, evolutionary cross-sections, showing the development of the Dabie Shan between the Sinian (late Proterozoic) and the middle Jurassic. For location see Fig. 10.

Chola Shan is the middle Triassic Jinsha Jiang suture (*XXI*).

In all of these suture zones, the Norian and later stratigraphic sections are represented by terrestrial sediments, despite the fact that *arc* magmatism in both the Chola Shan and in the Shaluli Shan Arcs continued generally into the Jurassic, and locally even into the early Cretaceous. Şengör (1984, 1985*a*) proposed that 'hidden subduction' beneath the overthickened flysch/*mélange* fill of the Songpan–Ganzi System may have been responsible for this peculiar observation. Thus, although marine conditions ended by the late Triassic in the entire Songpan–Ganzi system, subcutaneous 'hidden' subduction lasted until the early Cretaceous.

Although the structural style of the eastern margin of the Songpan–Ganzi System is vastly different from the south-west margin, characterized by enormous cuspate allochthons carrying thick Triassic clastics of the Bayanhar Group on to the Sichuan Basin, there too, hidden

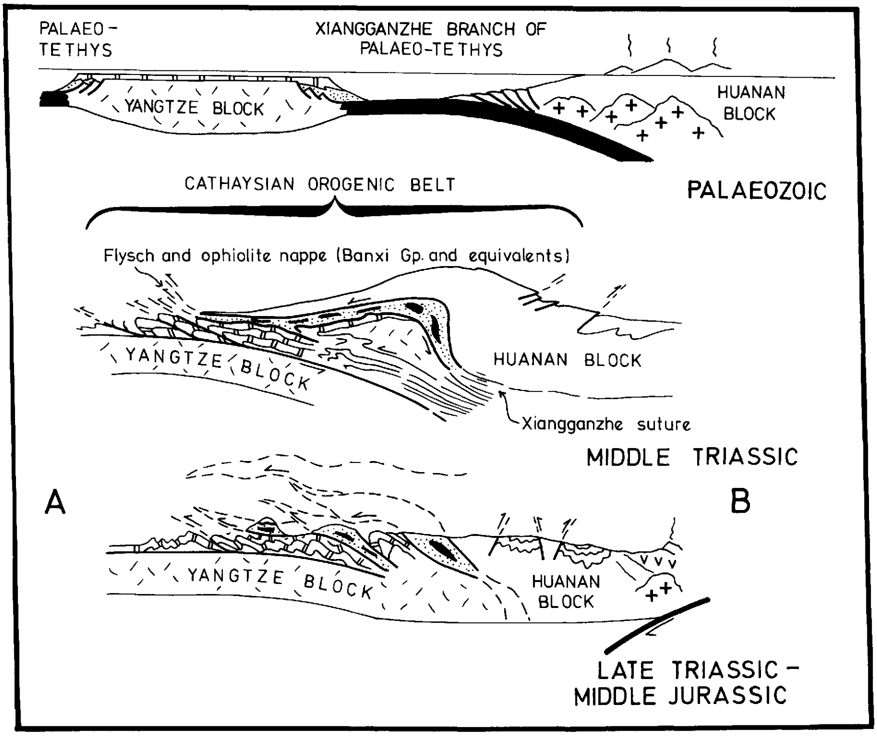


FIG. 12. Schematic, sequential, evolutionary cross-sections, showing the development of the Cathaysian orogenic belt between the late Palaeozoic and the middle Jurassic. For location see Fig. 10 (redrawn after Hsü *et al.* in press).

subduction may have played an important role in the Jurassic–early Cretaceous magmatic development of the Longmen Shan and Qionglai Shan, together forming the 'suture' *XXXIV*. But here a *line* of suturing between two continents is not present (except perhaps along the eastern margin of the Songpan Massif (*z*): Yang *et al.* 1986): 'suture' *XXXIV* is nothing more than a thrust-front, carrying the spilled-over flysch content of the constricted Songpan–Ganzi oceanic hole on to the Sichuan Basin (Figs 3 and 10).

Along the northern margin of the Songpan–Ganzi system, in the Anyemaqen Shan and Burhan Budai Shan, magmatic activity largely ended during the Triassic with very few and isolated Jurassic and Cretaceous plutons. Along the strike, westwards from the Anyemaqen Shan/Burhan Budai Shan 'suture' (*XX*) is the Southern Kuen-Lun suture (*XIX*). Along this suture, Şengör (1984) and Şengör and Hsü (1984), reviewed the available field evidence and concluded that collision probably took place by the early Jurassic.

Within the triangle defined by the Burhan Budai Shan (*XIX*) in the west, the Qin-Ling in the east (*XXXIII*) and the north-facing hairpin curve of the Litang 'suture' (*XXIII*), in the south, the Songpan–Ganzi System is filled with the enormous volume of the Triassic Bayanhar Group (which comprises dominantly turbiditic clastic rocks and some Permian deep-sea mudstones and radiolarian cherts, amounting to some five million cubic kilometres; Zhang 1981). The joint 1985 Royal Society–Academia Sinica Tibetan Geotraverse Team encountered these deposits between the Jinsha Jiang suture and the Kuen-Lun Pass on the Lhasa-Golmud highway. Along this stretch, they are made up of fine-grained, deep-water Triassic mudstones with graded wackes. The Geotraverse Team followed Şengör (1981, 1984) in interpreting these as parts of an accretionary prism made up of turbidites and contourites, but they also indicated that part of the sequence may have been deposited on a passive-margin continental rise. Because the Burhan Budai Shan/Anyemaqen Shan margin of Palaeo-Tethys was already a consuming margin, as judged from the abundant but very narrowly disposed late Palaeozoic arc-magmatic rocks,

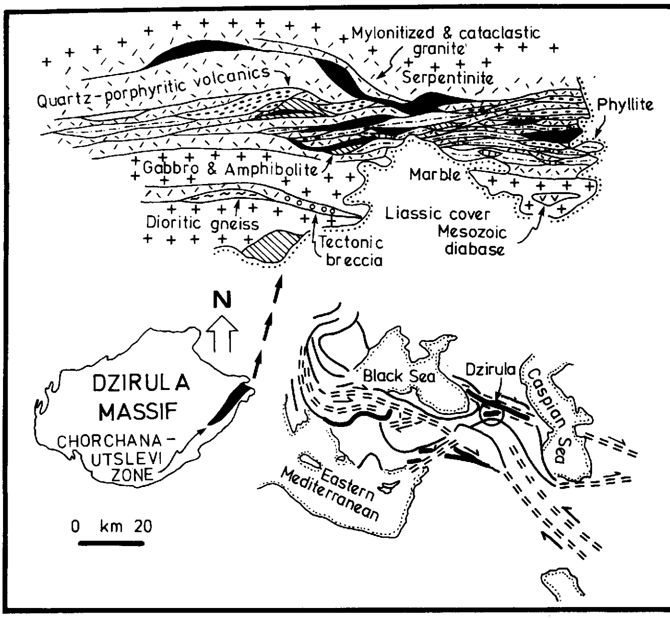


FIG. 13. Simplified geological map of the Chorchana–Utslevi shear zone of the Dzirula Massif. The location of the zone within the Dzirula Massif, and the location of the Dzirula area within the tectonic framework of the Tethysides, are shown in the two lower sketch maps. The lower right-hand map is taken from Fig. 22.

this conjecture seems unlikely. A forearc setting would be more likely for some of the Bayanhar lithologies.

A number of Cimmeride sutures are encountered under unconformable Liassic sequences in the western part of the orogenic zone. Although some of these may have already closed in the late Triassic, the available evidence seems to allow an intra-Liassic age for most of them. For only one of these, the Karakaya suture (*I*) in N. Turkey, related to a peri-Gondwanian back-arc basin of Palaeo-Tethys (Şengör and Yılmaz 1981; Şengör *et al.* 1984*a*), a late Triassic age seems at least locally well established (Tekeli 1981). Present data suggest that it probably began opening in the late Permian, at least in north-west Turkey, but may have been older farther east, near Erzincan (?Carboniferous) (Şengör *et al.* 1984*a*). Westwards this opening may have continued into the Maliac zone in Greece (Vergely 1984) or, in the early Triassic, into the Pindos–Budva Trough—probably also a Palaeo-Tethyan marginal basin (Şengör 1985*b*). Although it is not known whether the Karakaya suture extends any farther east, Y. Yılmaz (pers. comm. 1986) mapped sub-Liassic zones of intense deformation within the Malatya metamorphics (to the north-east of *22* in Fig. 3), thus extending the zone of sub-Liassic deformation previously known from the Menderes–Taurus block (known from the Lycian Nappes—north of *19* in Fig. 3—the Beyşehir autochthon—north of 18—and the Bolkardağ—north-west of *22*: Şengör *et al.* (1984*a*)) farther east. Şengör *et al.* (1984*a*) suggested that this deformation may represent the effects of the closure of the Karakaya marginal basin.

The Palaeo-Tethyan suture fragments in the Caucasian region all lie under unconformable Liassic sequences. The best outcrop of a Palaeo-Tethyan suture fragment in this region is in the Trans-Caucasian Dzirula Massif, where it is known as the Chorchana–Utslevi zone (V′; Fig. 13; Adamia and Belov 1984). It outcrops from under unconformable Lias, and consists of extremely sheared oceanic rocks and remnants of (?one) former magmatic arc. Their ages range from early Cambrian, through late Devonian to the late Palaeozoic (Adamia and Belov 1984). The present geometry of the Chorchana–Utslevi zone and its internal structures, suggests important amounts of strike-slip faulting following suturing. Both the process of suturing and the strike-slip faulting must have ended before the deposition of the overlying Liassic rocks. Coeval ophiolites also outcrop in the Khrami Massif along the Chochiani River gorge, within the

Artvin–Karabagh zone (*K* in Fig. 8a; Adamia and Belov 1984). Peive *et al.* (1980), Şengör (1984, 1985*a*), Şengör *et al.* (1984*a*), and Belov *et al.* (1986) argued that yet another, but cryptic, segment of the main Palaeo-Tethyan suture must be present in the Svanetia region of the south flank of the Greater Caucasus.

It is clear from the foregoing paragraphs that Palaeo-Tethyan suture segments in the Caucasian region occur in several different tectonic zones, whose formation post-dated Cimmeride tectonism. This suggests, as in Iran, that a formerly more regular pattern was later disrupted and led to suture tripling (from north to south: Svanetia, Dzirula, Khrami) in one cross-section in the Caucasus. We shall see in the following paragraph that this disruption of the Palaeo-Tethyan suture in the Caucasus was related to the same events that also disrupted the Palaeo-Tethyan suture in Iran.

Westwards from the Caucasus, the main Palaeo-Tethyan suture is younger. Eastwards, it is older in Turkmenia SSR and in northern Iran (along *VII*), but becomes younger again in Afghanistan and in the Northern Pamirs (along *VIII*). In these regions the Herat–Wanch–Akbaytal lineament (Stöcklin 1977) marks the main Palaeo-Tethyan suture (*VIII*), north of which was a large early Carboniferous to late Triassic subduction–accretion complex, that once formed the southern margin of Laurasia (Şengör 1984). The Lias witnessed final emergence, and in the northern parts of the Paropamisus and the Western Hindu Kush, plant-fossil-bearing Liassic clastics, equivalents of the Shemshak Formation in Iran, were laid down. Although the Triassic and older sediments were metamorphosed in these regions, younger sedimentary rocks were generally not metamorphosed (Shareq 1981).

North of the Herat–Wanch–Akbaytal lineament, Triassic to Jurassic granodiorites and tonalites (ages between 230 and 210 Ma) intruded the highly deformed late Palaeozoic and Triassic clastic and volcaniclastic rocks all along the Paropamisus (Berberian and Berberian 1981; Debon *et al.* 1986), and suggest an earliest Jurassic collision along the Herat–Wanch–Akbaytal lineament. Kravchenko (1979) describes a similar picture from the Northern Pamirs. In all these regions the first post-orogenic sedimentary rocks that unconformably overlie the suture are early Cretaceous red beds (Şengör 1984).

Palaeo-Tethyan suture zones younger than the Lias are known from the two extremities of the Cimmerides (not counting the Songpan–Ganzi System). In northern Turkey, the westerly continuation of the Palaeo-Tethyan suture (*IV*) from the Caucasus is represented by the ophiolites, deep-sea muds, and cherts, and the flysch plus shallow-water limestone knockers now making up the Küre Nappe (Şengör 1984; Şengör *et al.* 1984*a*; Yılmaz and Şengör 1985; Tüysüz 1986). Earlier workers assigned ages, ranging from questionable Permian to the earliest Dogger, to the lithologies of the Küre Nappe (see Şengör *et al.* 1980) a range confirmed by the TPAO Arama Grubu (1986). The TPAO Arama Grubu also found some late Carboniferous palynomorphs in the Küre lithologies (M. Aydın verbal comm. 1986; see also Yılmaz and Şengör 1985) thus suggesting that the age of the Küre Nappe oceanic assemblage might extend down into the Carboniferous.

Şengör *et al.* (1984*a*) interpreted the Küre Nappe as a large subduction–accretion complex, and correlated it with the diabase–phyllitoid association and the Lipacka flysch from the Strandja Mountains in Bulgaria (to the immediate south of *16* in Fig. 3; Chatalov 1979), with the Tauridian flysch in the Crimea (Beznosov *et al.* 1978; Kotansky 1978), and with the Nalbantian flysch in the North Dobrudja (Sandulescu 1978, 1980). Both in northern Turkey and in Bulgaria, sedimentation within the Küre Nappe largely ended by the end of the Liassic (the latest Triassic in Bulgaria) and it was tectonially covered by the Kırklareli, Istanbul, and the Bayburt Nappes, representing the northern margin of the Cimmerian continent before the (?middle) Dogger. This overthrusting also extensively disrupted an already complex internal structure within the Küre Nappe (see, for example Tüysüz 1985 for the details of this event).

Judging from the record of arc-type igneous activity within the Kırklareli, Istanbul, and the Bayburt Nappes, Şengör *et al.* (1984*a*) concluded that Palaeo-Tethyan subduction southwards under northern Turkey commenced during the late Carboniferous (Westphalian) and lasted until the early Jurassic. The latest representatives of this regime may have been the so-called Kastamonu granitoid belt, containing allegedly subduction-derived plutons (Boztuğ *et al.* 1984; Yılmaz and Boztuğ 1986). Both Boztuğ *et al.* (1984) and Yılmaz and Boztuğ (1986) considered the Kastamonu belt a product of *northward* subduction because, they observed, these plutons occurred to the north of some of the Küre Nappe ophiolitic rocks and high-grade metamorphics. However, because the Küre Nappe as a whole represents an accretionary complex, the tectonic position of the Kastamonu belt appears very similar to the late Triassic–early Jurassic magmatic arc in the Paropamisus and Western Hindu Kush; alternatively, it may be related to post-collisional thickening and partial melting of the Küre rocks, as originally suggested by Şengör *et al.* (1980). This latter interpretation is supported by the observation that the partial melting of subduction–accretion complexes commonly produces granodioritic magmas, as in the Chugach Mountains of southern Alaska (Hudson and Plafker 1982).

Where the main Palaeo-Tethyan suture in northern Turkey and in south-east Bulgaria extends to farther west is still debated. We here adhere to the scheme proposed in Şengör *et al.* (1984*a*), who suggested, following S. Dimitriadis' unpublished results, that the Therma–Volvi–Gomati ophiolites may mark the location of a pre-Tithonian Palaeo-Tethyan suture in Greece. They traced this suture farther north into eastern Serbia, into the region of Jastrebac, a suggestion followed here (northern part of suture *I* in Fig. 3).

III shows the Luncavita–Consul suture in the North Dobrudja. It represents a northerly-vergent thrust complex, bringing the Triassic Niculitel ophiolite on to the Tulcean Massif and uniting the latter with the Macin zone to the south of the suture. This suture probably closed in the early Jurassic. Şengör *et al.* (1984*a*) suggested that the whole of the North Dobrudja may have been a segment of the northern Turkish Cimmerides, fitting into the interruption of the suture *IV* (indicated by a dotted line). The North Dobrudja probably left this locality, and was transferred to its present position by strike-slip faulting during the earliest Cretaceous (Şengör *et al.* 1984*a*).

Farther north and west of eastern Serbia, indications of a Palaeo-Tethyan suture are reduced to slivers of undated ultramafic rocks, located within poorly-dated, high-grade metamorphic terranes within the Southern and Eastern Carpathians. Şengör (1984) discussed these in some detail, but it is now strongly doubted whether they really represent remnants of Palaeo-Tethys (M. Sandulescu, pers. comm. 1985). Thus, Greece and Yugoslavia seem, so far, to contain the westernmost Palaeo-Tethyan suture zones. In both of these countries, however, both the precise location and the age of Palaeo-Tethyan closure are still matters of dispute (e.g. Robertson and Dixon 1984; Vergely 1984; Mountrakis 1986; Şengör 1986*a*).

A probably early to mid-Jurassic Palaeo-Tethyan suture fragment in the eastern extremity of the Cimmerides is probably represented by the Serabang Formation in western Sarawak, and the equivalent rocks of immediately adjoining regions of Kalimantan (*XXXII*). These consist of strongly folded slaty flysch with chert and ophiolites (Haile 1976). Both Haile (1976) and Audley-Charles (1978) consider the Serabang Formation and its equivalents to be older than Jurassic. This general region is also characterized by early Mesozoic granite intrusions and coeval and younger felsic, intermediate, and mafic, lavas and tuffs (Fig. 5 and 6). Şengör (1984) argued that this region represented a pre-late Jurassic Palaeo-Tethyan suture. From here the suture probably goes through the West Borneo basement. Murphy (in press) suggested that a Carboniferous to late Triassic magmatic arc, extending along the hypothesized Serabang suture, is probably truncated at the late Cretaceous Meratus suture (*34*) which here defines the boundary between the Tethysides and the Circum-Pacific superorogenic complex (Murphy in press). Murphy (in press) correlates the Bornean Palaeo-Tethyan suture with the Nan Uttaraddit/Sra-Kaeo suture in Thailand. It seems, however, more likely that the latter suture continues directly into the Bentong–Raub line in the Malay Peninsula as argued previously, and that the Serabang/West Borneo suture is a prolongation of the Song Da suture in Vietnam (Şengör 1987*a*).

With the single exception of the Shilka suture in the Soviet Far East (*XLI*), the only Cretaceous sutures are related to the closure of the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean, the *Meso-Tethys* of Shvolman (1978) and Belov (1981). This substantial ocean split a considerable portion of the Cimmerian Continent into at least two halves, and may have caused the disintegration of the half lying to its north into two pieces in northern Tibet (*s′* and *s″* separated by *XXIV′*) (Montenat *et al.* 1986).

So far the most abundant information concerning this suture zone has come from the Afghan (*IX*) and the Soviet (*X*) sectors (for the most recent review of data from the Afghan and Tibetan parts see Montenat *et al.* 1986). Shvolman (1978) described the evolution of what he termed the Meso-Tethys, in terms of late Permian rifting, early (?to late) Triassic intracontinental stretching with abundant basaltic volcanism, later Triassic–Jurassic sea-floor spreading and Cretaceous closure along the Rushan–Pshart zone. A fairly complete ophiolite is described, involving a serpentinized harzburgite, containing dykes of gabbro and gabbro-diabase. The upper 200–300 metres of this peculiar gabbro–harzburgite complex are brecciated and contain, in addition to gabbros, also diorite and plagiogranites. This igneous section is overlain by sheared and differently metamorphosed calcareous schists and cherty greywackes and siltstones. A pile of variously spilitized tholeiitic lavas caps the metasedimentary section.

From Shvolman's (1978) descriptions one obtains the impression of a fossil fracture zone. Shvolman (1978) further argued that the tholeiite pile was the initial product of an ensimatic island arc, whose age reaches into the Cretaceous. We believe that this arc may have nucleated on a fracture zone.

The vergence of the present structure in the central and south-eastern Pamirs, along the boundary of which the Rushan–Pshart zone is located, is clearly northward (Shvolman 1978). From the location of most of the Cretaceous granites to the south of the Rushan–Pshart zone (see, for example, fig. 1 in Desio 1977) one obtains the impression of a north-facing orogenic polarity which also fits the structural vergence. The presence of Permian diabases and andesitic basalts on *both* sides of the Rushan–Pshart zone (Shvolman 1978) suggests that the Rushan–Pshart ocean may have opened as a back-arc basin disrupting a continental margin arc during the late Permian, much like the Karakaya marginal basin in northern Turkey (*II*; see also Fig. 20a).

Both westwards and eastwards the Rushan–Pshart zone is cut and offset by large, Cainozoic, strike-slip faults (Fig. 3). The Jurm, Zebak–Mun'ya, and Harirud Faults offset it south-westwards by some 400 km (Shvolman 1977, fig. 52; Bazhenov and Burtman 1986; fig. 1) where it continues westwards as the Waser (or the Farah–Rud) suture (*IX*). The Waser suture (*IX*) is also a north-vergent structure (Blaise *et al.* 1978). Within it, the fossils recovered from the volcano-sedimentary carapace of the Waras ophiolite indicate a middle to late Triassic spreading age. Although no direct evidence exists for the timing of initial rifting of the Waser ocean in Afghanistan, analogy with the Rushan–Pshart zone makes it likely that this too began rifting during the late Permian or the early Triassic at the latest. The closure of the Waser ocean took place during the early Cretaceous with a northerly vergence. Cretaceous granodiorites and arc volcanics indicate the southerly subduction of its floor, but Debon *et al.* (1986) argue that the late Cretaceous ages abundant in these magmatic rocks may indicate a later age of closure here than hitherto considered (Fig. 7), although its suture is unconformably covered by *Orbitolina*-bearing limestones of Aptian–Albian age (Blaise *et al.* 1978).

In Tibet, the Banggong Co–Nu Jiang suture (*XXIV*) has been studied mainly in its eastern half. In its western part, observations are sparse, but Norin (1976, 1979) identified the south-east continuation of the Rushan–Pshart zone in the Aghil–east Loqzung Mountains of north-west Tibet. The Aghil Ranges are offset to the south-east from the Rushan–Pshart zone by the Karakorum Fault for about 100 km or less. The Loqzung Mountains are characterized by Permian limestone, quartzites, quartz sandstones, and shales of marine origin, with which mafic porphyries seem to be associated (Norin 1946). Towards the south, north of the Banggong Co, the possible along-strike continuations of Norin's 'porphyries' are seen as ophiolites. In the best-studied segment of the Banggong Co–Nu Jiang suture zone, around the Donqiao–Gyanco area of eastern Xizang (to the north-east of *t''* in Fig. 3), Upper Palaeozoic low-grade metamorphic grey limestone, quartzites, and shales, similar to the Loqzung facies, are seen to be tectonically interleaved with well-developed ophiolites. Both the Upper Palaeozoic metasediments and the ophiolites were thrust southwards above a mid- to late Jurassic flysch and all were then unconformably overlain by Aptian–Albian red clastic rocks and volcanics (Girardeau *et al.* 1984*a*). The Banggong Co–Nu Jiang suture was closed before the *Orbitolina*-bearing late Cretaceous limestones were deposited upon it in a wide region (Bally *et al.* 1980).

Recently, Srimal (1986) suggested that the Banggong Co suture should be correlated westwards with the Shyok suture north of the Ladakh batholith. This suggestion is based on comparisons between poorly-dated ophiolite slivers from the northern Saltoro Range and Banggong Co, and does not take the geology of the Central and south-eastern Pamirs into account. We shall return to the question of the age and significance of the northern Saltoro ophiolites in the following paragraphs.

Farther east, the suture zone follows the course of the Nu Jiang and turns abruptly southwards around the eastern syntaxis of the Himalaya (Fig. 3). From there it goes to the Mandalay region (to the north-west of *XXIX*), where it is doubled back on to itself and on to the southern continuation of the Indus–Yarlung Zangbo suture (*31*) along a north-south right-lateral strike-slip fault, whose offset is about 400 km in the Mandalay region (Mitchell in press, Curray *et al.* 1982).

From Burma southwards, we follow Şengör's (1984) correlation of the earliest Cretaceous (?latest Jurassic) deformation, metamorphism, and granitic intrusions in the basement of the eastern Sumatra Basin (Katili 1976) with the

Mandalay zone. Murphy (in press) recently also adopted this interpretation.

With the discussion of the long suture zone of the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay/east Sumatra oceanic realm we have come to the end of our discussion of the known Cimmeride sutures. We should perhaps mention three other sutures in this connection, although neither their age, nor their relationship to the Cimmeride evolution is known with any certainty, nor are they generally considered as Cimmeride sutures by the majority of the workers active in this field. One of these is the Shiquan He suture (*XXV*) reported by Mei *et al.* (1981) and Baud (1985). Baud (1985) encountered ophiolites along the Shiquan He during a traverse from Ladakh to Yarkand. Previously, Norin (1976) had indicated the presence of what he called the Upper Shyok(?)–Shaksgam 'rift zone' (Norin called all major sutures 'rifts' including the one along the Indus) and implied a major zone of disturbance characterized by a distinctive black shale facies. This zone may possibly be identical with, and represent the north-west continuation of, the Shiquan He suture. How the Shiquan He–Shaksgam zone may continue into the Pamirs to the west is unknown, but we will mention the possibility of its being represented by the late Palaeozoic–early Mesozoic black slate troughs of the south-eastern Pamirs, south of the Rushan–Pshart suture zone (see fig. 1 in Norin 1976). This possibility is indicated schematically in Fig. 3 by the conjectured suture fragment *b* (for 'black slates') to the south-east of the Rushan–Pshart zone (*X*). If our tentative correlation across the Karakorum fault is correct, then the Shiquan He suture should have closed before the Bajocian (see Norin 1976, 1979). Mei *et al.* (1981, p. 551), however, indicated that 'Cretaceous continental basalt-andesite-rhyolite formations are developed . . . on the . . . Nganglung Kangri Range' north of the Shiquan He ophiolites (see Mei *et al.* 1981, fig. 1), implying that the Shiquan He ocean was still open during the Cretaceous. This is the interpretation preferred by Mei *et al.* (1981), although, curiously, Srimal (1986) considers the Shiquan He ophiolites to be late Palaeozoic, for which there is no evidence. The arc-type magmatic rocks of the Nganglung Kangri Range may, on the other hand, belong to a south-dipping subduction zone associated with the Banggong Co–Nu Jiang ocean, and thus be unrelated to the Shiquan He suture.

A. Baud (verbal comm. 1985) and Srimal (1986) suggest that the Shiquan He suture zone may continue westward into the southern Tibetan Lhasa block, and in fact may have separated the Trans-Himalayan late Mesozoic–early Cainozoic magmatic arc from regions lying to the north of it. They indicate that the Xainxa ultramafic rocks located within the Lhasa block (near 31°N lat. and 88°E long.; Girardeau *et al.* 1985) may be the eastern prolongation of the Shiquan He suture zone. A. Gansser (verbal comm. 1985) further suggested in support of this view, that the entire Lhasa block may have consisted of four separate blocks (t', t'', t''' and t''''). However, the recent field observations of the 1985 Royal Society–Academia Sinica Tibetan Geotraverse Team seem to suggest that the ophiolitic outcrops within the Lhasa block actually represent remnants of a single, northerly-derived sheet of ophiolite, that roots into the Banggong Co–Nu Jiang suture zone, and that became disrupted by post-obduction deformation of the Lhasa block itself. The geological and geochemical studies of the Lhasa block ophiolites by Tang and Wang (1986) also corroborate this conclusion.

Chang *et al.* (in press) recently reported total sediment thicknesses in excess of 15 km from within the Qangtang block (*s*). This observation, suggesting at least substantial crustal thinning, may indicate the presence of a major suture zone within the Qangtang block, when we combine it with earlier reports of glaucophane schists from within the same block (Hennig 1915). Suture *XXIV′* in Fig. 3 was drawn by connecting the observation points of Chang *et al.* (in press) and Hennig (1915). This suture probably closed by the end of the Triassic, because middle Jurassic sediments lie above the Triassic rocks across an important unconformity around its southern end (Chang *et al.* in press).

The third possibly Cimmeride suture is the south-west Karakorum suture (*XXVI*), reported by M. P. Searle (verbal comm. 1985). Neither the age nor the along-strike continuations of this zone are known in any detail, but the northern Saltoro ophiolites, reported to be of late Palaeozoic age by Srimal (1986), represent the direct east–south-east prolongation of the metamorphosed harzburgites, serpentinized dunites, and garnet pyroxenites reported by Searle *et al.* (1987) and Bertrand and Kienast (1987) from within the Karakorum metamorphics south of the Axial Batholith, between the Masherbrum Glacier in the south-east and the Panmak Valley in the north-west.

Bertrand and Kienast (1987) indicate the presence of a large amount of metagreywackes within the Karakorum metamorphics, in which the ultramafic rock slivers are found. The probably Carboniferous 'flysch-like' successions, reported from along the Yasin and Ishkuman Valleys in the Hindu Raj region by Casnedi (1980) may be correlative deposits, and may imply the presence of a small late Palaeozoic oceanic trough in this region, that may have stretched from the present Saltoro range to the Hindu Raj region. When this trough closed is unknown, but its closure and the metamorphism of the Karakorum metamorphics must have predated the Mango Gusar granite intrusion (36.4 ± 1 Ma U–Pb zircon age) (Searle *et al.* 1987). Casnedi (1980) pointed out that shallow-water Permian sediments followed the ?Carboniferous deep-sea deposits and implied the possible presence of a pre-Permian tectonism here. On the basis of regional considerations, we do not believe that the age of closure of this trough was much younger than the early Cretaceous.

Alpide sutures

Compared with the broad, eastward-widening net of the Cimmeride sutures, the better-known Alpide sutures are much simpler, commonly consisting of single strands for long stretches, with the notable exception of a braid of sutures between Yugoslavia and Iran, which is still small compared with the Cimmeride net. In the following account we briefly review their geology and emphasize only the more controversial points.

Mesozoic Alpide sutures

In the entire Alpide orogenic system there are very few sutures that closed terminally during the Mesozoic Era, the majority having formed during the early Cainozoic. The earliest Alpide suture is the southern and eastern sections (the Nish–Trojan suture: Hsü *et al.* 1977) of the Circum-Moesian suture zone (*10*) that closed by south-dipping subduction beneath the Rhodope–Pontide fragment (*f*) in the Albian (Hsü *et al.* 1977, Burchfiel 1980).

The next youngest Alpide suture is the so-called Northern Suture (*29N*) north of the Kohistan island arc (*m'*) and its eastern equivalent, the Shyok suture (*30N*) in northern Ladakh (*t''''*). Petterson and Windley (1985) have shown that the Kohistan segment of this suture must have closed between about 100 Ma and 75 Ma ago, that is, roughly between the Albian and Campanian.

Geological relationships and the timing of events are similar in Ladakh (*t''''*). We agree with Petterson and Windley (1985) that both Kohistan and Ladakh used to be parts of a single ensimatic arc system that formed just offshore of the southern margin of Eurasia about 100 Ma ago. This united arc collided with Eurasia during the middle Cretaceous and became a continental-margin arc.

Two distinct suture belts of probably pre-Maastrichtian age exist in Turkey and in northern Greece. Şengör and Yılmaz (1981) have argued that the Peonias–Intra-Pontide suture (*13*) must be of pre-Maastrichtian age, because the associated arc along the Black Sea coast terminated its activity at this time. However, continuing arc volcanism farther south, and the fact that the oldest sediments unconformably covering the deformed rocks of the suture in Turkey are of Lutetian age (Tokay 1973), may be indicative of a later closure.

In south-east Turkey, along the Assyride suture (not exposed; covered by the products of *21*), the Bitlis–Pötürge–Malatya–Keban portion of the Menderes–Taurus block (the eastern end of the African promontory, *e*) is believed to have collided with the northern margin of Arabia in pre-Maastrichtian time (Şengör and Yılmaz 1981). Widespread ophiolite obduction southward on to the Arabian platform was associated with this collision. A nearly coeval collision was postulated by Gealey (1977) to obduct the giant Semail ophiolite nappe in Oman (*27*). Recent geochemical studies by Searle *et al.* (1980) and Pearce *et al.* (1981) confirm Gealey's (1977) arc collision hypothesis.

Both of these obduction events took place along the southernmost continental margin of the Alpides in the Middle East facing an ocean that had already begun rifting in the Permian (in the Zagros) and the early Triassic (southern Turkey). Until the middle Triassic, intracontinental extension continued in both places, followed by sea-floor spreading in Oman, and also probably in the Zagros in the late Triassic. In southern Turkey and around the eastern Mediterranean (southern branch of Neo-Tethys) the geological record of the continental margins suggests that sea-floor spreading probably commenced in the Jurassic. A side-branch of the eastern Mediterranean, in the form of a small embayment, formed the Pamphylian Basin in south-west Turkey in the middle to late Triassic (Şengör and Yılmaz 1981) by rifting and rotation of a small fragment of the Menderes–Taurus block, the Alanya Massif (*e'*).

In Sumatra, Cameron *et al.* (1980) interpreted the record of the late Mesozoic Woyla Group as indicating the latest Jurassic–early Cretaceous opening, and late Cretaceous (probably between 110 and 85 Ma ago) compressional obliteration of a marginal basin, whose suture is today represented by the Woyla Group (*33*). This suture may continue eastwards into the island of Java and then turn to the north-east, to re-emerge eventually in the Meratus Mountains of south-east Borneo (*34*), where ophiolite obduction occurred between the Cenomanian and Turonian (Murphy in press).

Cretaceous collisions have been proposed for other Alpide sutures as well (e.g. Tapponnier 1977, for the Alps (*8*), Bergougnan and Fourquin 1980, for the İzmir–Ankara–Erzincan suture (*14* and *17*), Okay 1984, for the İzmir–Ankara suture (*14*), Berberian and King 1981 for the Zagros suture (*25*)), but most of these are based on overinterpretation of the available data, or result from erroneous conceptions of tectonic processes, or are based on negative evidence. Some of these authors have later retracted their original erroneous interpretations (e.g. Berberian *et al.* 1982). One reason why so many workers have favoured late Cretaceous continental collisions along the Alpides has been that the late Cretaceous was a time of very widespread ophiolite obduction and blueschist metamorphism—from Corsica to Oman. More recently, Searle (1983) has argued that late Cretaceous ophiolite obduction also must have taken place in the Himalaya, on to the north-west margin of the Indian Shield in the Zanskar Range (immediately south-east of *30* in Fig. 3) but there is little evidence to support this conjecture (see Şengör 1987*b*).

It is difficult to prove that all of the ophiolite obduction events in the Alpides were not associated with continental collisions (cf. Ben-Avraham *et al.* 1982), but so far no evidence has been found to show that the late Jurassic–early Cretaceous ophiolites in Yugoslavia and Greece (Spray *et al.* 1984), late Cretaceous ophiolites in Corsica (Gruppo di Lavoro sulle Ofioliti Mediterranee 1977), in Turkey (Şengör and Yılmaz 1981), and in the Zagrides (Ricou 1976) were associated with any collision. They all appear to have been pre-terminal obduction events (cf. Dewey 1976), mostly caused by the interaction of oceanic fracture zones with passive continental margins.

Cainozoic Alpide sutures

It is possible to divide the Cainozoic suturing events in the Alpides into two large classes: Palaeogene and Neogene sutures.

The Vardar zone and the İzmir–Ankara suture appear to have closed during the Palaeocene (Burchfiel 1980; Şengör and Yılmaz 1981; Vergely 1984). The easterly continuations of both the Intra-Pontide suture and the İzmir–Ankara suture (*13* and *17* respectively) closed later, during the Lutetian and Priabonian. Along all of these sutures, ocean opening took place mainly during the Hettangian–Sinemurian interval (Görür *et al.* 1983), with the possible exception of the İzmir–Ankara and the Vardar sutures, along which rifting was probably earlier—in the Triassic.

In north-east Turkey, in a number of places it is difficult to draw a line of demarcation between the Karakaya and the İzmir–Ankara sutures. In such places one obtains the impression that the Karakaya ocean did not completely close, small remnants of it having survived between sutured segments, and that these remnants were later incorporated into the İzmir–Ankara ocean.

The small Pamphylian oceanic basin in south-west Turkey began closing in the late Cretaceous and closed terminally during the latest Palaeocene–Eocene by north-vergent thrusting along the Antalya suture (*19*; Şengör and Yılmaz 1981; Ozgül 1985).

To the south of Ankara (immediately south of *11* in Fig. 3) the Karakaya ocean was probably in direct communication with the Inner Tauride ocean that began opening in the early Triassic (Özgül 1984). This Triassic opening appears to have been a local intracontinental rifting event only, because regions lying to its south—the entire Menderes–Taurus block—seem to have been affected by the closure of the Karakaya Basin to the north (Şengör *et al.* 1984*a*, Y. Yılmaz, pers. comm. 1986). The Inner Tauride ocean closed during the middle Eocene (Görür *et al.* 1984).

Another latest Palaeocene–earliest Eocene closure was that of the Waziristan (*28*)/Main Mantle Thrust (*29S*)/Indus–Yarlung Zangbo (*30S* and *31*) closure that welded India on to Eurasia (Fuchs 1984; Petterson and Windley 1985). Collision along this suture seems to have progressed from west to east during the Eocene, as a result of which the entire Indian subcontinent rotated in an anticlockwise fashion for about 15° (Klootwijk 1981).

The timing of opening of the Indus–Yarlung Zangbo ocean, i.e. the Himalayan Neo-Tethys, is fairly well-established by observations on both the Lhasa block and on the northern margin of the Indian Shield. In the south-west part of the Lhasa block (t''''), an overall shallowing and local emergence is seen above a 500-metre thick Lower Permian limestone horizon. Near the Indus–Yarlung Zangbo suture zone the same overall shallowing is observed. During the middle to late Triassic a mixed volcanism with andesites and basalts is seen, accompanied by clastic sedimentation and normal faulting (Bally *et al.* 1980).

On the Indian side, a rapid subsidence during the earliest Triassic is accompanied by extensive normal faulting and fissuring (Bassoullet *et al.* 1978) in Ladakh (along *30S*). Widespread Triassic turbidites flank almost the entire northern margin of the Indian Shield, and indicate the establishment of a north-facing passive continental margin. It thus seems clear that the Himalayan Neo-Tethys came into being by the middle to late Triassic at the latest.

The consumption of the floor of Neo-Tethys was already under way by north-dipping subduction under the Lhasa block at large, during the Portlandian (late Jurassic). Initially, this subduction probably began within the oceanic floor of Neo-Tethys. (Giardeau *et al.* 1984*b*). Chang *et al.* (in press) pointed out the presence of small patches of mainly Lower and Middle Cretaceous and partly Upper Jurassic andesite–dacite–keratophyre associations, forming part of the basement of the later Trans-Himalayan magmatic arc. These volcanic rocks show the presence of calc-alkaline volcanism during the late Jurassic–early Cretaceous interval along the southern margin of the Lhasa block. Thus, models involving a late Jurassic rifting of this block away from north-west Australia (e.g. Rowley *et al.* 1986) do not appear to be compatible with data from it.

The onset of the consumption of the Neo-Tethyan ocean floor in the Himalayan segment, post-dates the obduction of the Banggong Co–Nu Jiang ophiolites by about 20 Ma only (obduction *c.*170 Ma, onset of magmatism *c.*150 Ma), although the major, widespread Trans-Himalayan plutonism did not begin until about 110 Ma ago (Xu *et al.* 1985).

Another latest Palaeocene to earliest Eocene collision was the encounter of the African Promontory with the European south-facing continental margin in the Alps, along the Alpine suture (*8*). Ongoing controversy still characterizes discussions on the original palaeogeography, polarity of subduction zone(s), and the timing of events in the Alps. We here adopt the palaeogeographical scheme proposed by Trümpy (1971, 1975, 1980; Trümpy and Haccard 1969) and refined by Lemoine and Trümpy (1987), because other models (e.g. Dercourt *et al.* 1986) do not account for the available observations nearly as fully as does Trümpy's scheme. Late Cretaceous southerly-vergent structures in the Pennine zone (e.g. Milnes 1978) and in the Austro-Alpine nappes of eastern Switzerland (e.g. Şengör 1982*a*) seem to be related to reactivation of north-dipping normal faults of the southern continental margin of the Alpine ocean in a dominantly north-facing subduction geometry, as recently confirmed by the presence of Turonian to earliest Coniacian trench deposits in the Walsertal zone along the boundary of the Penninic and Austroalpine units (Winkler and Bernoulli 1986). However, the possibility of a south-facing subduction zone along the southern margin of the Briançonnais mini-continent (fragment contained between the sutures *7* and *8* where they meet in Fig. 3), as originally proposed by Hsü and Schlanger (1971), remains open.

The age of opening of the Alpine ocean, especially the Piedmont–eastern Valais segments (northernmost part of *7* and the part of *8* lying to the east of its junction with *7*) is fairly well established, that is: Hettangian–Sinemurian rifting (with a Rhaetian prelude) and subsidence (Bernoulli and Jenkyns 1974; Lemoine and Trümpy 1987) followed by the middle Jurassic onset of ocean-floor spreading (Isler and Pantiç 1980; Lemoine and Trümpy 1987).

Although the age of rifting remains remarkably uniform on both sides of the Alps proper—both westwards through the Apennines and into the Tellian Atlas/Riff/Betic sutures (*7*, *6*, *3*, and *2* respectively) and eastwards into the Pieniny Klippen belt (*9*), the times of ocean closure vary greatly. Moreover, ocean-floor spreading took place west of the Alps, in the Apennines (Bigazzi *et al.* 1972), probably in the Pyrenees (Boillot and Capdevila 1977) and also possibly in the Betic chain (*ophiolites bétiques*? in Wildi 1983). Between Sicily and Gibralter no ophiolites have yet been confirmed. The ultramafic nappes of Ronda in southern Spain (Wildi 1983) and Beni Bouchera in the Riff (Reuber *et al.* 1982) are subcontinental ultramafic rocks exposed by extreme extension during the middle Mesozoic. In the Apennines, ocean closure occurred mainly in the Oligocene (Amodio-Morelli *et al.* 1976, Kligfield 1979), whereas along the westernmost Mediterranean chains it was later, mainly in the early to middle Miocene (Wildi 1983). As Şengör (1984) pointed out, because the opening of the Alpine ocean (and its westerly continuation into the Apennines and beyond) was a part of the opening history of the Central Atlantic ocean, it is more correct to consider these mountain belts as extra-Tethyan and parts of a separate system: the *Atlantides*, products of the closure of a part of the Atlantic Ocean (see Fig. 1). However, because the extant data are not sufficient to separate the Tethysides and the Atlantides with confidence, we continue to include the latter into the Alpides.

The Pyrenees represent the evolution of a small ocean basin that opened in the early Cretaceous and closed in the middle Eocene (Boillot and Capdevila 1977). An earlier, Triassic rifting nearly coincident with the future Pyrenean axis, failed to produce oceanic lithosphere, and became dormant by the Jurassic (Ziegler 1982). We do not follow the models portraying the Pyrenees as a dominantly strike-slip-related orogen not involving a Wilson cycle (e.g. Choukroune *et al.* 1973, Daignieres *et al.* 1982), simply because the Triassic salt lying on the Landais marginal plateau at the southern apex of the Bay of Biscay (Valéry *et al.* 1971) makes it geometrically impossible to slip the Iberian Peninsula into its present place solely along a strike-slip fault, without involving a significant extension both in the Bay of Biscay and in the Pyrenean region. Moreover, convergent structures north of the Iberian Peninsula (Montadert *et al.* 1974), showing the former existence of a north-facing subduction zone, support the model of Boillot and Capdevila (1977), and suggest that the Pyrenean orogen is the product of the subductive removal of a small ocean and subsequent continental collision.

Eastwards, in the Carpathians, the Neo-Tethyan suture becomes progressively younger around the Carpathian bend east of the Rhodope–Pontide fragment, in the prolongation east and south-east of the Pieniny Klippen belt zone (*9*), and then in the so-called 'external flysch trough' of the eastern Carpathians (Sandulescu 1975; Burchfiel and Royden 1982).

Collision along the southerly prolongation of the Pieniny Klippen Belt, in the Apuşeni Mountains (*11*), and in the Vardar zone of eastern Serbia (*12*) was nearly coeval with that in the Alps (Burchfiel 1980, Sandulescu 1980, Grubiç 1980). This more 'internal zone' extends southward to connect with the Vardar zone *sensu lato* in northern Greece (*12*, *13*).

Thus an inspection of the suturing episodes around the African promontory in the circum-Adriatic regions reveals a progression of suturing in both directions away from the Alps. As Caire (1977) has shown, this progressive closure of the westernmost Tethyan–Atlantic branch resulted in an accentuation of the curvature of the orogenic belts around the African promontory in the course of the Tertiary, a phenomenon further intensified in the west by the opening of the western Mediterranean basins (Cohen 1980).

Another event that narrowed the African promontory in the south, and accentuated the Z-shape of the Carpathian–Hellenide segment of the Alpides was the latest Eocene–Oligocene closure by westward-thrusting of the Pindos–Budva Trough. Deformation since this time has migrated farther west in Greece and Yugoslavia (Richter *et al.* 1978) and active folding and thrusting now characterize the coastal regions of these countries facing the Adriatic Sea (McKenzie 1972, 1978).

In the Balkan Ranges, the Srednogorie tectonic province represents a former marginal basin that began opening during the Cenomanian–Senonian transition, behind the north-dipping Peonias–Intra-Pontide suture zone (Antonijevič *et al.* 1974, Aiello *et al.* 1977, Hsü *et al.* 1977). This marginal basin was originally the western continuation of the Black Sea marginal basin (Görür, in press), but closed along the suture *16* during the Eocene, with a northerly vergence (Hsü *et al.* 1977).

Along the southern slope of the Greater Caucasus the Slate-Diabase zone ocean (*24*) also closed in the Eocene by south-vergent thrusting (Khain 1975).

Two other suture belts of Oligocene age, in addition to the Apennines and the Pindos–Budva zone, are the Sistan suture in eastern Iran (*26*; Tirrul *et al.* 1983) and the Indo-Burman suture (*32*) along the north-eastern border of the Indian Shield (Brunnschweiler 1966). The Sistan suture ('East Iranian Flysch Trough' of Stöcklin 1974, and Berberian and King 1981) represents an ocean that probably opened during the early Mesozoic as a continuation of the Waser ocean in Afghanistan, and closed by east-dipping subduction between the Campanian and the Oligocene (Tirrul *et al.* 1983).

We close this survey of the Tethyside sutures, with the Neogene sutures of the Alpides which are found at two extremities of the Tethyside superorogenic system around the loop of Gibraltar (*2* and *4*), in the island of Timor in the extreme south-east of the Tethysides (*35*), and around the Arabian Peninsula (*21*,*25*).

Ricou (1973), Şengör and Kidd (1979), Şengör and Yılmaz (1981) and Berberian *et al.* (1982) documented the late-middle Miocene to Pliocene ocean closure along the Bitlis–Zagros suture zones (*21*,*25*), that finally welded Arabia to Eurasia.

Cimmeride and Alpide sutures: a comparison

As the earlier discussion shows, convergence along many of the Alpide sutures is still in progress, and thus not only the Alpide orogen is being enlarged continuously in such regions, but also its fore- and hinterland areas are being disrupted extensively by germanotype structures

(Fig. 2). The Alpides thus represent a still-active system. In fact, in a number of places, Neogene sutures pass along the strike into still-active subduction zones (e.g. *21* westwards into the Cyprus subduction zone, *25* and *28* into the Makran subduction zone, Indo-Burman sutures and the Timor suture into the Andaman–Sumatra–Java subduction zone).

Mini-oceans opened as a consequence of the destruction of the Neo-Tethys, mostly as back-arc basins and/or pull-apart structures along large strike-slip belts, such as the western Mediterranean basins (Cohen 1980), the Black Sea (Görür in press), and the southern Caspian Depression (Berberian 1983) and these also still survive within the Alpide edifice.

None of the Cimmeride sutures today continue into active subduction zones. The mini-oceans generated by the Cimmeride tectonics no longer remain within the Cimmeride edifice, with the possible exception of the Tarim, Junggar, and Qaidam Basins (Hsü in press). This shows that, in contrast to the Alpides, the Cimmerides represent a fossil orogen, whose ruins served as a foundation to the later construction of the Alpides. Like the Alpides, the Cimmerides also generated a vast field of germanotype cratonic structures in Eurasia (Fig. 2) that developed between the late Permian and the early Cretaceous, with the greatest intensity and areal extent being in the late Triassic–Jurassic (Şengör 1984, 1985*a*). Many of these structures also terminated their activity by the middle Cretaceous, but some remained active, being inherited by the Alpides (Şengör 1984).

A few of the sutures within the Tethyside edifice represent oceans that *opened as a part of the Cimmeride evolution, but closed during the growth of the Alpides*. For example, the Pindos–Budva Trough opened as a Palaeo-Tethyan marginal basin, possibly as the westerly continuation of the Karakaya mini-ocean (Şengör 1985*a*, 1986*a*), but closed in the latest Eocene–Oligocene, as a result of suture progradation from the Almopias–İzmir–Ankara suture zone. Similarly, the Sistan ocean in eastern Iran opened as a part of the Waser–Banggong Co–Nu Jiang ocean and closed in the Oligocene by the westward expulsion of the Helmand block. In all these cases we classified the associated sutures as Alpide, because their *closure* was a part of the Alpide tectonics, independent of any Cimmeride influence. After all, significant stretches of Neo-Tethys itself—in eastern Turkey, Iran, Pamirs, south-east Asia—seem to have opened initially as Palaeo-Tethyan back-arc basins, and only during the Jurassic did their united descendants acquire an independent life of their own as the Neo-Tethys (Şengör 1979*a*, 1984). Such genetic relationships, more than spatial coincidence, induced Şengör (1984) to combine the Cimmerides and the Alpides under one common system, the Tethyside superorogenic complex. Many of the blocks contained within the Tethyside orogenic collage are bounded by both Cimmeride and Alpide suture segments. In order to understand the evolution of these sutures and the growth of the collage, it is necessary to know the original provenance and the route of migration of these Tethyside blocks. In the next section we briefly review the relevant information to establish these two parameters.

Blocks of the Tethyside collage

Fig. 3 shows all of the known Tethyside blocks ('block' is here used in the sense of a primary collage component, as defined by Şengör, 1987, and in press). Four major lines of evidence are currently available to establish the original provenance and the path of migration of the Tethyside blocks. In the order in which they are commonly employed, these are palaeobiogeography, palaeoclimatology, stratigraphy and structural evolution, and palaeomagnetism.

Palaeobiogeographical and palaeoclimatological data from the Tethyside blocks

In Figs 14 through 17 we have plotted the distribution of various palaeobiogeographical and related palaeoclimatological indicators in and around the Tethysides, to see to what palaeobiogeographical and palaeoclimatic provinces the various Tethyside blocks might have belonged at various time intervals. The data we report are not exhaustive, but were chosen mainly to identify Laurasia versus Gondwana Land affinities. For more comprehensive summaries of palaeobiogeographical and palaeoclimatological data the reader is refered to Ziegler *et al.* (1979), Parrish *et al.* (1982), Parrish and Curtis (1982), Buffetaut *et al.* (1984), Nakazawa and Dickins (1985), Rowley *et al.* (1985), and the papers in this volume.

Late Carboniferous through early Permian

Fig. 14 displays the distribution of late Carboniferous to early Permian floras, cold-water faunas, and evidence for glaciation. Evidence for glaciation both from south of the Indus–Yarlung Zangbo suture in India (*31*) and west of

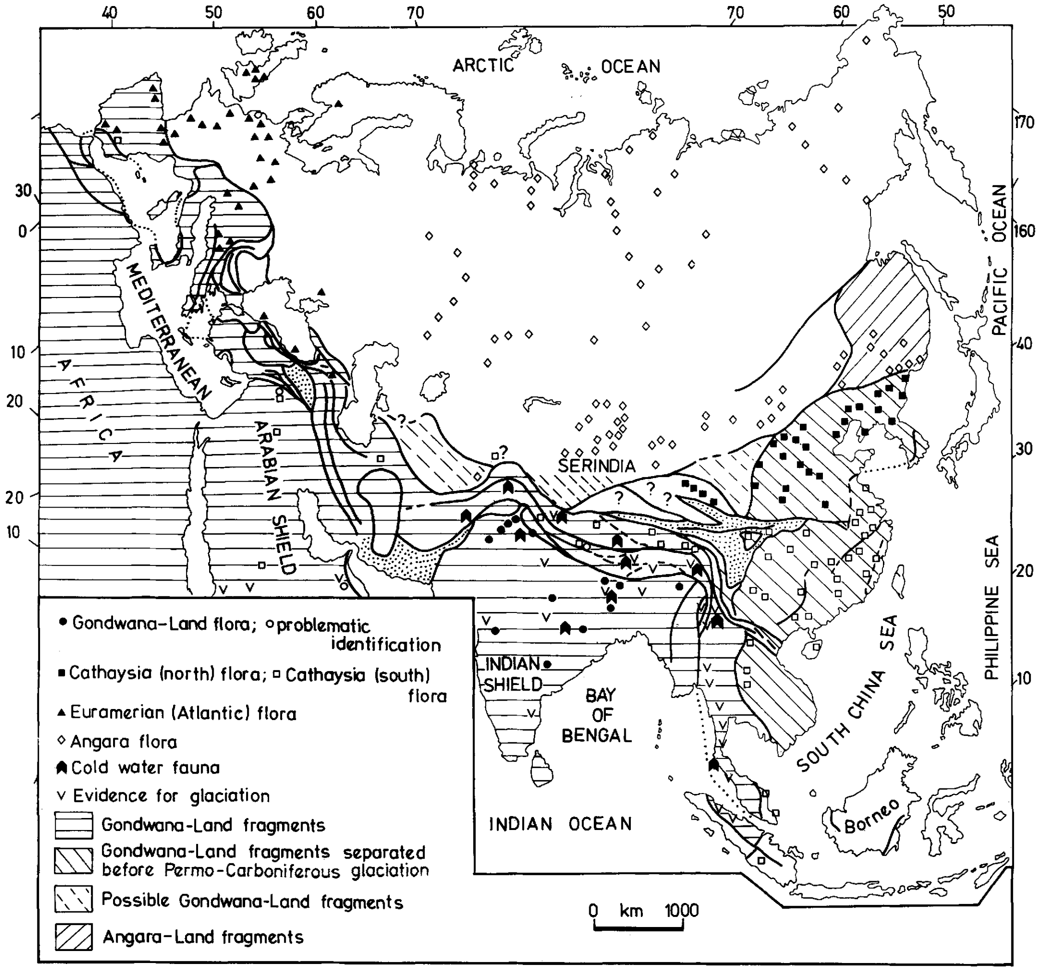


FIG. 14. Distribution of late Carboniferous–early Permian floras, cold-water faunas, and traces of glaciation in and around the Tethysides. Modified from Şengör (1987*a*), mainly using Kerey (1984), Belov *et al.* (1986), Cao (1986), and W. Chaloner (written comm. 1986).

the Zagros–Oman (*25,27*) suture on the Arabian Peninsula, and from the north of the Neo-Tethyan sutures (Fig. 14) belong to the 'direct evidence' category, as classified by Crowell (1983). Although the glacial nature of the Damxung–Linzhu 'tillites' located in the eastern part of the Lhasa block (in the northern part of *t‴*) was questioned by Allègre *et al.* (1984), there is at least one locality reported earlier from farther east in the same block (near Baxoi, in the extreme east part of *t″*), that has unquestionable Middle and Upper Carboniferous tilloids (Chen 1981). Later, Lin (1983) reported Upper Carboniferous to Lower Permian glacio-marine sediments from the Xainza County near the central part of the Lhasa block. Most recently, Chang *et al.* (1986) not only confirmed the glacio-marine tilloid nature of the Upper Carboniferous Damxung mixtite, but they also reported the presence of striated pebbles from at least two localities on the Lhasa block, plus dropstones. This, combined with earlier reports, now puts the Gondwanian affinity of the Lhasa block, and its close proximity to glaciated terrains, as late as the early Permian, beyond dispute. Similarly, the presence of both glacio-marine deposits in north-west Qangtang (*s′*) (Norin 1946, 1976), and the discovery of rich *Eurydesma* faunas in three separate localities within the same block (Liu and Cui 1983), indicate that the *western part* of the Qangtang block (*s′*) was also a part of Gondwanaland until at least the early Permian. Evidence from the eastern Qangtang block (*s″*) indicates that this block has a Cathaysian flora.

No evidence for glaciation has yet been reported from this block, and Chang *et al.* (1986) conclude that it was probably no longer part of Gondwanaland by the Permian.

Recently, Altermann (1986) questioned the tilloid aspect of the Carboniferous to early Permian pebbly mudstones of Thailand, located on the Sibumasu block (s'''). His arguments, however, are based on unsupported estimates of the age span of these deposits, insufficient attention to relationships between pebble (and boulder) roundness and lithology, misunderstanding of the significance of the included diamonds, and neglect of the palaeontological evidence. A general review of these deposits by Stauffer and Mantajit (1981), and a detailed analysis of the outcrops in the Langkawi islands by Stauffer and Lee (in press) strongly support the glacio-marine origin of these deposits. Moreover, Cao (1986) recently reported the presence of striated pebbles and the cold-water forms *Stepanoviella* and *Eurydesma*, in glacio-marine deposits of late Carboniferous age from Baoshan, located in the northernmost extremity of the Sibumasu block (s''') in western Yunnan.

From the data shown in Fig. 14 the following conclusions emerge. Until mid-Permian times, almost the entire Asiatic part of the Cimmerian Continent (s', t', t'', t''', m) constituted a part of Gondwanaland. In three localities, Euramerian late Carboniferous floras have been found on the Cimmerian continent. Two of these are in northern Turkey, both from the Istanbul and Bayburt nappes, and one is from the Khrami Massif in Transcaucasia. Belov *et al.* (1986) argue that this may be because of the westward narrowing of the Palaeo-Tethyan triangle, enabling some floras to 'cross the suture'. The 'North China Foldbelt' block was already a part of Laurasia in the late Palaeozoic, as was the Turan block, whereas the other Chinese blocks (q, z, w, y at large), Annamia (x), and eastern Malaya, were away from both Gondwana Land and Laurasia, and were probably located somewhere near the equatorial latitudes. A Cathaysian flora occurs on the Arabian Peninsula (Gondwana Land!) and in Iran. The Iranian occurrence was probably similar to those on the Arabian Peninsula, i.e. located on Gondwana Land.

Because both fossils of land plants and marine macro-organisms are not as frequently encountered in the Tethysides as those of micro-organisms, we have also plotted the distribution of certain foraminifera within and around the Tethysides, for the late Permian, early Jurassic, and late Jurassic to Neocomian times (Figs 15, 16, and 17). Similar studies recently undertaken by others (e.g. Ross and Ross 1981; Ross 1982*a*, *b*; Ishii *et al.* 1985; Okimura *et al.* 1985, for the Permian; Cherchi and Schroeder 1973; Schroeder *et al.* 1978; Moullade *et al.* 1980, 1985; Bassoullet and Bergougnan 1981; Pélissié *et al.* 1982; and Bassoullet *et al.* 1985) have contributed to the palaeobiogeographical analysis of the Tethyan domain mainly in two aspects. Firstly, diagnostic forms for palaeobiogeographical provinces have been identified from among cosmopolitan taxa. Secondly, their distributions have been plotted on, and checked against, continental reconstructions.

Many ecological factors, such as temperature, salinity, depth, or pH of water, control the distribution of benthonic foraminifera. A change in one or more of these, might at times form ecological barriers. These barriers may be generated by the presence of oceans between different centres of populations, and by ecological changes within these oceans and their peripheral shelf seas, caused by plate motions. The benthonic foraminifera constitute excellent tools in palaeobiogeographical studies to check continental reconstructions.

The three maps we present here (Figs 15, 16, and 17) have been distilled from a large data base. Both the details of the individual fossil localities and the relevant references will appear in a separate palaeontological paper, including a discussion on different biozonation schemes of the three selected time intervals (Altıner, in prep.), to which we refer the interested reader.

Late Permian

From mid-Permian times onwards, the glacially influenced sediments and biotas disappear, and the biogeographically significant marine faunas gain in importance for purposes of continental reconstruction. In the evolution of the late Permian foraminifera, two main trends are observed towards the end of this epoch. The large fusulines (*Eopolydiexodina*, *Neoschwagerina*, *Yabeina*, *Lepidolina*, etc.) disappeared before the upper limit of the 'Midian' (Leven 1981; see Altıner 1984 for a discussion) and the number of taxa of the smaller foraminifera (*Shanita*, *Paraglobivalvulina*, *Paradagmarita*, *Colaniella*, etc.) notably increased.

In Murghabian time, *Neoschwagerina* spread over the entire Tethyan domain, with the exception of central, south-central, and eastern Turkey, large parts of the Zagros shelf, Oman, and also possibly western Yunnan in China, Burma, and western Thailand. Although not shown in Fig. 15, the genus *Verbeekina* seems to follow very much the distribution pattern of *Neoschwagerina*, albeit not as frequently encountered. Both of these genera are therefore very largely confined to Cimmerian continent, the Chinese blocks south and east of the sutures *XIX*, *XX*, *XXXIX*, *XV* and *XVI*, and to a few spots on the northern Gondwanian shelf as the southern and central parts of the African promontory (*e*), the Abadeh area in the Zagros, and the Tibetan Himalaya. Neoschwagerinids thus seem to have come in contact with Laurasia during the late Carboniferous collision of the North China block with this continent. Otherwise, they seem confined entirely to Tethyside blocks derived from Gondwana Land, and to Gondwana Land itself.

Eopolydiexodina, usually accompanied by *Rugososchwagerina* (latter not shown in Fig. 15) is known from Yugoslavia, Turkey, Transcaucasia, Afghanistan, the Darvas region in Turan, Burma, and western Thailand. This form has not been encountered in China, Indo-China east of the Nan-Uttaradit/Bentong Raub suture (*XXVII*, *XXXI*) and Japan, making it highly significant for palaeobiogeographical analysis. *Eopolydiexodina* thus seems confined only to the Cimmerian Continent, and to the north and north-east shelf of the Arabian Peninsula (Gondwana Land), with the exception of Darvas in Turan, whose tectonic position is still uncertain.

Among the smaller foraminifera of the late Permian, *Dagmarita* (not shown in Fig. 15) is found to be a cosmopolitan form and *Abadahella* spreads over the regions also occupied by the genus *Neoschwagerina*.

During the 'Midian', *Lepidolina* spread over Japan, southern China, and Indo-China, and thus formed an important province in eastern Asia, including only those blocks of Gondwana Land origin that had left the supercontinent before the Permian. The only exception is the

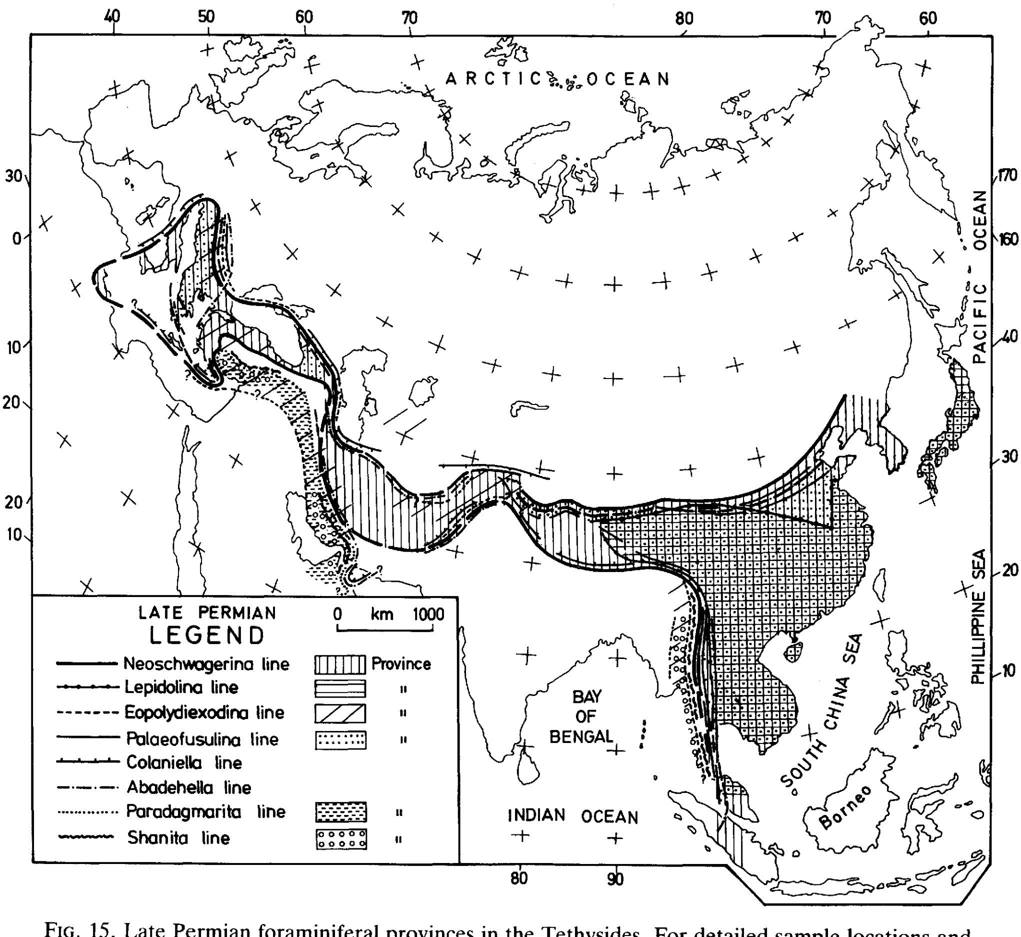


FIG. 15. Late Permian foraminiferal provinces in the Tethysides. For detailed sample locations and sources see Altıner (in prep.).

Lhasa block where *Lepidolina* also occurs. *Shanita* is also a good palaeobiogeographical marker, being found only in southern and south-east Turkey, in the Zagros shelf, in Oman, and in western Yunnan, Burma, and Thailand. It seems also confined to the Zagros margin of Gondwanaland and to the future Neo-Tethyan margin of the Cimmerian Continent.

The latest two stages of the late Permian epoch, the Djulfian and Dorashamian, are palaeobiogeographically characterized by three genera, namely: *Palaeofusulina*, *Colaniella*, and *Paradagmarita*, which are also excellent index fossils. The aberrant *Palaeofusulina* is a typical eastern Asian form, being abundant in Japan, China, and Indo-China, but it is also found in the western Tethyan domain (Caucasia, Greece, and Yugoslavia). This genus is absent in Afghanistan and probably also in Iran. It probably had a very narrow path of migration along the northern margin of the Cimmerian continent. Interestingly, the places where it is absent are marked by south-dipping subduction north of the Cimmerian continent, where quiet platform sedimentation was disturbed, as in northern Turkey, Iran, and the Pamirs.

The genus *Colaniella* has a large distribution and commonly follows the distribution pattern of *Neoschwagerina*. *Paradagmarita*, on the other hand, has a very restricted distribution and is found only in southern and south-east Turkey, and along the Zagros shelf and in Oman.

In summary, the genus *Neoschwagerina* spreads over vast areas of the Tethyan domain but is generally limited to the Cimmerian continent, except in China and Indo-China. In the central Mediterranean region, where the Cimmerian continent was hinged to Pangaea, and in the Tethyan Himalaya, it spills over on to Gondwana Land. Genera such as *Yabeina*, *Abadahella*, and *Colaniella* follow the same distribution. Two well-defined provinces are formed by the genera *Paradagmarita* and *Lepidolina*, in the Middle East and Far East, respectively. The distribution of *Shanita* corresponds with that of *Paradagmarita* in the Middle East, whereas in Burma it occurs unaccompanied by this form where *Neoschwagerina* and *Lepidolina* are also absent. *Eopolydiexodina* and *Lepidolina* never spatially coincide. In the Far East, *Eopolydiexodina* occurs together with *Shanita*. Finally, the genus *Palaeofusulina* seems to have spread over into the western Tethyan regions from its main region of distribution in the Far East, via a narrow path possibly along the northern margin of the Cimmerian continent.

Early Jurassic

In the early Jurassic, the most spectacular development in the complex agglutinated foraminifera is seen in the populations of the genus *Orbitopsella* (*O. primaeva*, *O. praecursor*, *O. dubari*), and the other phylogenetically related form, *Cyclorbitopsella*. In the Pliensbachian–Domerian, the genus *Orbitopsella* spread over the Arabian Platform, Morocco (High and Middle Atlas), Tunisia, northern Egypt, Israel, Syria, south-east Turkey, the Zagros shelf, and Oman (Fig. 16). In Europe, its distribution follows the Betic chains of southern Spain, Balearic islands, Italy, Dinarides in Yugoslavia, Greece west of the Vardar zone (*12*) and southern Turkey. In Central and East Asia, the data on the distribution of *Orbitopsella* and the related genus *Cyclorbitopsella* are fragmentary. They have been reported from the Lhasa block (*t*), from the Markam region in the extreme eastern end of the eastern part of the Qangtang block (*s*″), and from western Thailand. This pattern of distribution shows that, although in the Mediterranean and Middle Eastern regions *Orbitopsella* was confined to regions south of the northern branch of Neo-Tethys, in Asia it was found both on independent blocks within the Tethyan domain (such as the Lhasa block) and along the Laurasian southern margin in Markam and western Thailand. This Asiatic anomaly is now difficult to explain.

The palaeobiogeographical distributions of many taxa follow the limits of the *Orbitopsella* province. These are *Mayncina termieri*, *Pseudocyclammina liassica*, *Lituosepta* (*L. recorensis*, *L. compressa*), *Pseudopfenderina butterlini*, and *Haurania*. The last genus however has been recorded in western France and in north-west Tibet, indicating an anomalous excursion in these regions, far into Laurasia outside the *Orbitopsella* province (Fig. 16).

Late Jurassic to Neocomian

Many complex and agglutinated late Jurassic larger foraminifera, such as *Kurnubia palastiniensis*, *Parurgonia caelinensis*, *Kilianina lata-rahonensis*, *Labyrinthina mirabilis*, *Alveosepta jaccardi*, and *Anchispirocyclina lusitanica*, show a cosmopolitan nature. These taxa spread over a large band of terrain, limited in the north along a line from northern France to Kopet Dagh, and in the south from Morocco to Senegal. The occurrence of late Jurassic foraminifera in Central and East Asia has been reported from only a few localities, and the

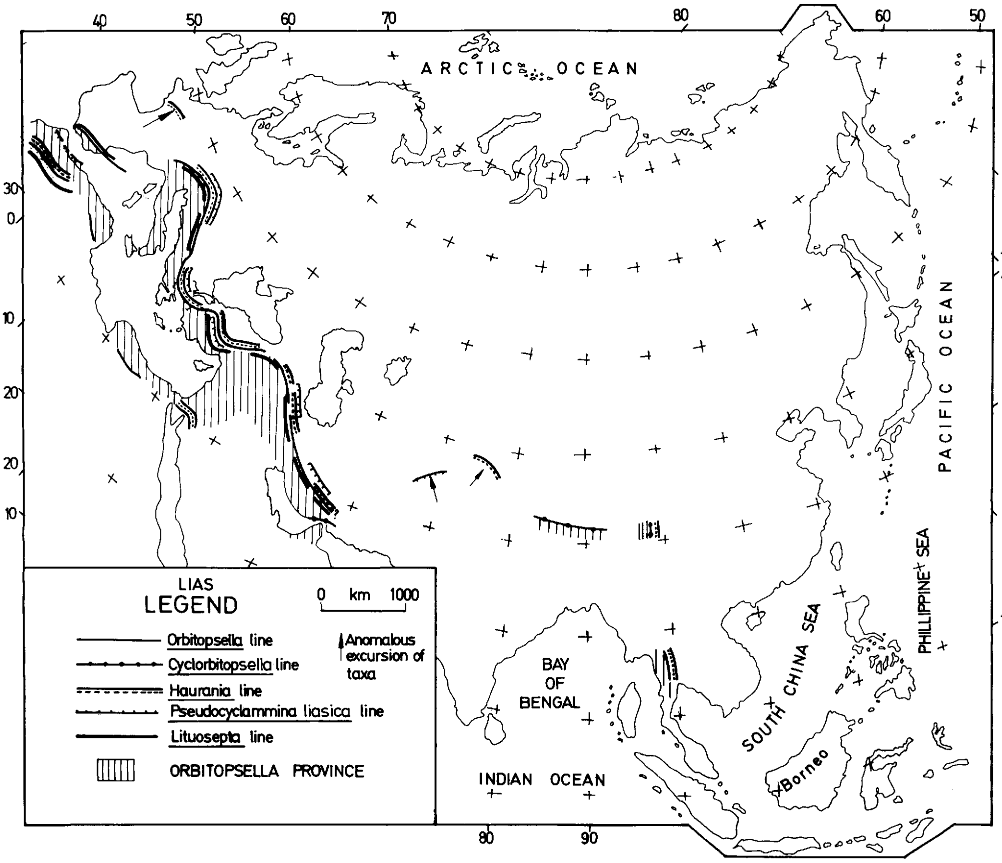


FIG. 16. Early Jurassic (Lias) foraminiferal provinces in the Tethysides. For detailed sample locations and sources see Altıner (in prep.).

easternmost occurrence is known from the Philippines (Fig. 17).

Within the broader realm of the late Jurassic larger foraminifera it is possible, however, to define a narrower and more useful palaeobiogeographical zone. Algeria, Tunisia, Italy, Yugoslavia, and Greece south-west of the Vardar zone (*12*), southern and south-eastern Turkey, Iraq, Syria, Israel, Egypt, Saudi Arabia, Qatar, Oman, and the Zagros Shelf are regions where the genus *Kurnubia* exhibits a tremendous plexus in its evolution. The various steps in the evolution from *K. palastiniensis* to *K. wellingsi*, with the other related form '*Conicokurnubia*', are not observed in the *Kurnubia* populations found in the regions to the north of this narrower biospace. This *Kurnubia* plexus province (Fig. 17) coincides with regions lying to the south of the northern branch of Neo-Tethys and the Zagros/Oman suture zone (*6*, *7*, *8*, *11*, *12*, *13*, *14*, *17*, *25*) and seems to be delimited rather sharply by this major oceanic space.

In the Neocomian, several species of agglutinated foraminifera described from Western Europe constitute important but small palaeobiogeographical provinces, and their tectonic usefulness is thus limited (Fig. 17). One exception to this is *Orbitolinopsis capuensis*, distributed in northern Algeria, Tunisia, Italy (mainly in the Apennines), Yugoslavia, Greece west of the Vardar zone (*12*) and the Menderes–Taurus block (Fig. 17). This form thus appears to have been confined to the African Promontory and to its extension in Turkey—the Menderes–Taurus block. In the Oran Meseta (*b*) the *Orbitolinopsis capuensis* and the *Choffatella pyrenaica* provinces overlap (Fig. 17).

Many other Neocomian larger foraminifera, such as *Protopeneroplis trochoangulata*, *Pseudotextulariella salevensis*, and the long-ranging forms recorded also from the Upper Jurassic or above the Neocomian (e.g. *Conicospirillina*, *Trocholina alpina*, *Trocholina elongata*, *Nautiloculina cretacea*) are cosmopoli-

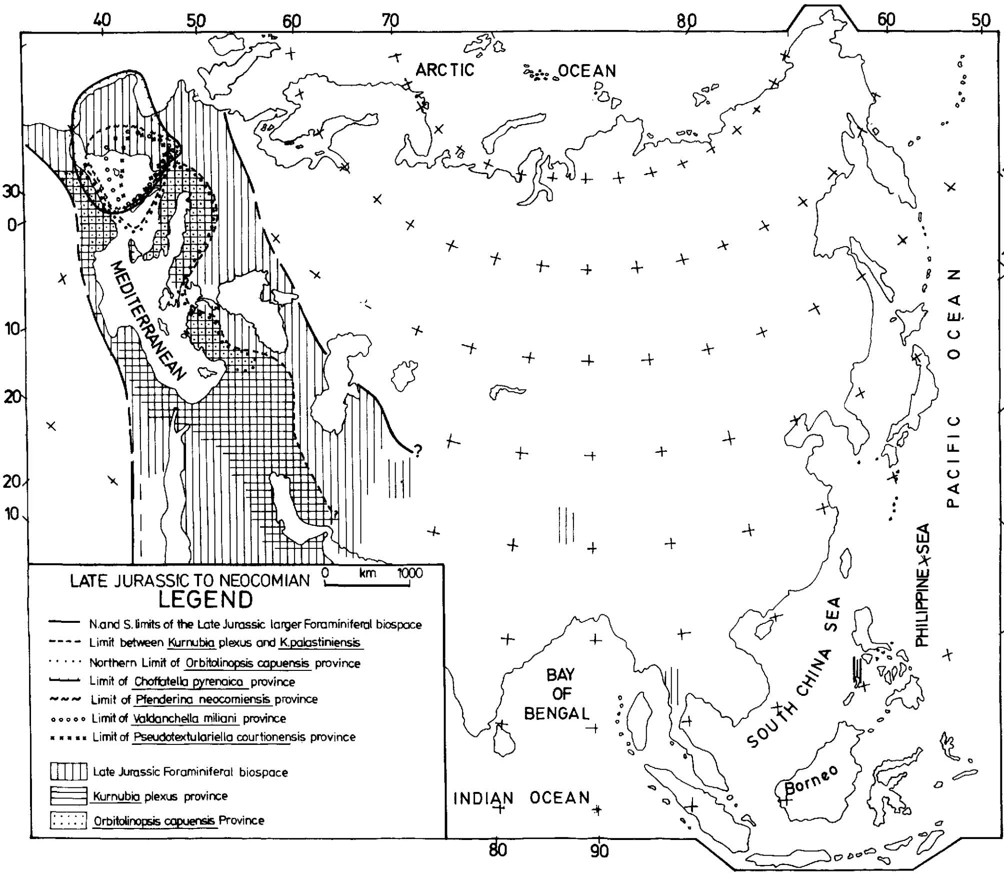


FIG. 17. Late Jurassic to Neocomian foraminiferal provinces in the Tethysides. For detailed sample locations and sources see Altıner (in prep.).

tan forms, and their palaeobiogeographical distributions coincide more or less with those of the late Jurassic forms.

The palaeobiogeography of certain foraminiferal genera presented here, shows the following features from a tectonic viewpoint. In the Permian, *Neoschwagerina* and the associated genus *Verbeekina* define the components of the Tethyside collage, except for blocks *a*, *b*, *c*, *d*, *n*, *o*, *p*, and *r*, and spill over on to Gondwanaland only in a few places along its northern shelf south of the Neo-Tethyan sutures. These forms did not everywhere cross the future Neo-Tethyan sutures in Asia east of Turkey. In these regions the southern boundary of the *Neoschwagerina* and *Verbeekina* province is mostly coincident with, or close to, the future Neo-Tethyan rift axis. This province crosses the future northern Turkish Neo-Tethyan sutures, and those in the Apennines, indiscriminately, corroborating their post-late Permian age.

The sharp S-shaped *Neoschwagerina* province boundary in western Turkey and Cyprus may be a result of lack of detailed information from the central Turkish regions where the facies are not exposed, but it may also be indicative of a left-lateral motion in post-Permian times.

Both the early Jurassic *Orbitopsella* and the late Jurassic *Kurnubia* plexus provinces are delimited to the north by the northern branch of Neo-Tethys and the Zagros suture, except in Asia where *Orbitopsella* has been reported both from an intra-Tethyan block (Lhasa) and from the southern margin of Laurasia (Markam, Thailand). Neither of these two provinces seem to be affected by the presence of the southern branch of Neo-Tethys (now represented by the sutures *19*, *20*, the buried Assyrian suture under *21*, and the eastern Mediterranean itself), which formed a boundary only for the *Orbitolinopsis capuensis* province.

Palaeobiogeographical and palaeoclimatological data of late Carboniferous to early Permian age thus largely show that the Cimmerian Continent almost entirely constituted the northern and north-east margin of Gondwana Land at this time, with the exception of the small eastern Qangtang fragment. The North China (*q*), Yangtze (*w*), Huanan (*y*), and Annamia (*x*) major blocks, and the eastern Qangtang (*s″*) fragment plus the Shaluli Shan (*u*) and the Chola Shan (*v*) fragments, were somewhere in the tropics. They all share a Cathaysian flora, but the North China block is distinct in having a northern Cathaysian flora, as opposed to a southern Cathaysian flora common to all the others. Both the Turan block and the North China Foldbelt (*r*) were by this time parts of Laurasia. Although both northern Turkey (*f*) and Transcaucasia (north-western parts of *i* and *j*) have Euramerian floras, we consider them to be parts of Gondwana Land, because all other data become unintelligible otherwise. These localities were not very far away from other Euramerian localities such as Yugoslavia.

In the later Permian, the foraminifera may suggest the beginning of separation of the Cimmerian continent from northern Gondwana Land in Asia east of Turkey, and may underline the unity of the Cimmerian continent, again with the exception of eastern Qangtang (*s″*).

In the Mesozoic, the Neo-Tethyan northern branch in the Mediterranean region, and the Zagros ocean, emerge as important barriers to foraminiferal dispersion, the southern branch of Neo-Tethys in the present eastern Mediterranean being important only for the *Orbitolinopsis capuensis* province. Both in the early Jurassic and in the late Jurassic–Neocomian interval, the Saharan Atlas was a southern boundary for foraminiferal distribution in North Africa.

At no time have the Pyrenean (*1*), Pindos–Budva (*15*), and the Sevan–Akera–Qaradagh (*23*) sutures been important barriers in agreement with the small size of the oceans involved (e.g. Pindos–Budva), or their short life-span (e.g. Pyrenees).

The distribution of Mesozoic land quadrupeds within the Tethysides has been a difficult topic to study, because of the dominantly marine nature of most of the available record. Recently, however, Buffetaut (1981) reported a number of Mesozoic quadrupeds from Thailand indicating a clear 'Laurasian' affinity since the latest Triassic.

Palaeomagnetic data from the Tethyside blocks

Independent support for most of the conclusions drawn in the last section is supplied by palaeomagnetic studies which we do not review here. For a good overview of the available sources of information the reader may consult Şengör (1987*a*). Newer data from the Yangtze block have been recently reported by Huang *et al.* (1986) and Opdyke *et al.* (1986).

Basement geology of the Tethyside blocks

Once palaeobiogeography, palaeoclimatology, and palaeomagnetic studies inform us in what latitudinal range, in what orientation, and possibly on what side of a major continent our object of interest must have been located, it is once more regional geology that will enable us to make a more detailed reconstruction by matching geological markers across suture zones.

Within the Tethyside orogenic collage, one of the best markers is the now-disrupted Pan-African orogenic system, making up the basement of many of the Tethyside blocks. Şengör (1986*a*) pointed out that the Pan-African 'event', stretching between about 900 Ma and 450 Ma (Kröner 1984), was responsible for the assembly of Gondwana Land—age of orogeny becoming progressively younger towards the margins of the supercontinent, as seen especially well in the Arabian Peninsula, India, and Australia. He further argued, that in the Asiatic Tethysides, the Tethyside rifting and orogenic events were commonly the first significant tectonic events to follow the terminal Pan-African events. Şengör *et al.* (1984*b*) showed that in Turkey, for example, events previously called 'Caledonian' actually constituted a continuation and the termination of the Pan-African orogenic phenomena. Although, unfortunately, still called 'Caledonian', the same is true for early Palaeozoic events in the Alps, (*e*: Frisch *et al.* 1984) in Iran (*i*, *j*, and *k*: Davoudzadeh *et al.* 1986), in Afghanistan (*m*: Montenat *et al.* 1981), in the Lhasa block (*t*: Chang *et al.* in press), in Annamia (*x*: Workman 1977) and in the Huanan block (*y*: Shui *et al.* 1986). Fig. 18 shows the distribution of possible Pan-African-related structures and their broad age groups. If we could reconstruct the original Pan-African geology of these fragments we might be able to reconstruct the Pan-African orogenic system, and thus at least the northern and north-east part of Gondwana Land, which would be equivalent to reconstructing the pre-dispersal geometry of the majority of the Tethyside collage components. Unfortunately, the existing data base is not nearly sufficient to do this, most of the available information from the Asiatic Tethyside blocks being merely in the form of ages of deformation and/or ages of unconformable successions lying on the Pan-African terrains. Less frequently, grade of metamorphism is crudely indicated, but the kind of metamorphism, structural geometry, type of associated magmatism, or the protoliths of the metamorphic rocks, are only rarely reported. With the present information, the best that one can do is to point out an overall trend of younging towards the periphery of Gondwana Land, from about 900 Ma to Ordovician (*c.*450 Ma), and to expect that when the Tethyside blocks are returned to their original positions in Gondwana Land this geometry would be reproduced.

As Fig. 18 shows, within the various blocks making up the Cimmerian continent, this geometry is preserved even today, with the exception of Iran (*i*, *j*, and *k*), south-east (*e*) and north-west Turkey (*f*). In southern China and Indochina this logic would demand that the Yangtze block (*w*) and the coastal block (*y″*) be placed farther towards the interior of Gondwana Land than the Huanan block (*y* and *y′*) and Annamia (*x*). This speculation is supported by independent Cambrian palaeomagnetic data from the Yangtze block, and stratigraphic data from both the Yangtze and the Huanan blocks (Lin *et al.* 1985*b*, fig. 6). The exceptional situation in Iran arises because the younger (early Palaeozoic) Pan-African terrain of the Sanandaj–Sirjan zone (*i*) is now located between older (latest Precambrian) Pan-African terrains of Arabia and Central Iran (*j*). This anomaly is automatically resolved to produce the expected

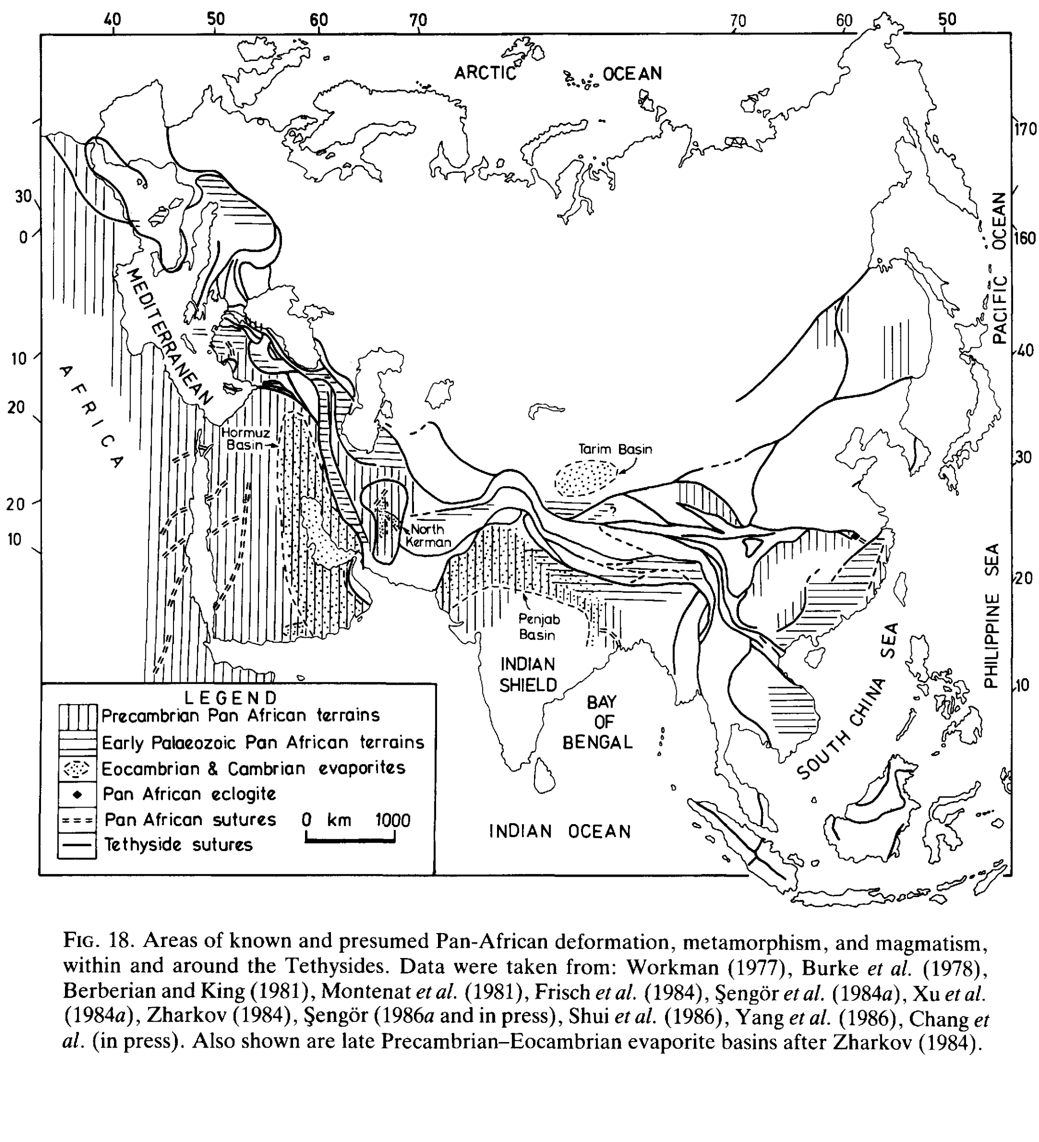


FIG. 18. Areas of known and presumed Pan-African deformation, metamorphism, and magmatism, within and around the Tethysides. Data were taken from: Workman (1977), Burke *et al.* (1978), Berberian and King (1981), Montenat *et al.* (1981), Frisch *et al.* (1984), Şengör *et al.* (1984*a*), Xu *et al.* (1984*a*), Zharkov (1984), Şengör (1986*a* and in press), Shui *et al.* (1986), Yang *et al.* (1986), Chang *et al.* (in press). Also shown are late Precambrian–Eocambrian evaporite basins after Zharkov (1984).

'younging outwards' geometry when the disrupted Palaeo-Tethyan suture and arc fragments in Iran are reconstructed to their original geometry (see next paragraph). The same also applies to north-west Turkey.

Fig. 18 also shows the latest Precambrian–early Cambrian evaporite basins within and near the Tethysides (Zharkov 1984). These basins formed synchronously with, or shortly after, the terminal Pan-African collisions in the present central Tethyside region. The Hormuz evaporite basin is delimited to the north-east by the Zagros suture, but correlative evaporites reappear on the 'Iranian' side of the suture across the Sanandaj–Sirjan zone, near Nayband and north of Kerman (Fig. 18). When the Central Iranian Microcontinent is rotated back to its pre-late Triassic orientation, as mentioned previously (Davoudzadeh *et al.* 1981), the north Kerman Cambrian evaporites become juxtaposed against the Sanandaj–Sirjan zone, which thus becomes the only obstacle separating them from the main Hormuz basin—a further point underlining the previously discussed anomalous position of the Sanandaj–Sirjan zone, and supporting Central Iran's Gondwanian origin.

In Fig. 18 we also show areas that were deformed during the late Precambrian in the North China Foldbelt region (*r*) and in the Qinghai Nanshan (*o*), but no other evidence exists to show that these events, although coeval, may also have been parts of the Pan-African evolution.

Although none of the palaeobiogeographical or geological data we discuss are relevant to the original provenance of the North China block, Lin *et al.* (1985*b*) present both Cambrian stratigraphic and palaeomagnetic data to show that it too belonged to Gondwana Land, as originally speculated by Şengör (1984) on the basis of a broad similarity of early Palaeozoic stratigraphy.

Fig. 14 summarizes the results of this section. The evolutionary scheme outlined in the next section is based on the data reviewed in this and the previous sections. As a detailed analysis of the Phanerozoic facies distribution in the Tethysides and a similarly detailed structural compilation are not yet available, the following evolutionary scheme, summarized in Figs 19 and 20, is only a very generalized and oversimplified picture. Numerous complexities implicit in the evolution of the Tethyside collage could not be dealt with, both because of the chosen scale and lack of detailed information. Only as an illustration of what these complications are, and how they might affect our reconstructions, we will briefly review the late Palaeozoic to late Cretaceous palaeotectonic evolution of Iran and surrounding regions.

Evolution of the Tethyside collage

Figures 19a to 19f display the essential steps in the assembly of the Tethyside orogenic collage. The Tethyan domain is defined with reference to the Permo-Triassic Pangaea, and therefore no 'Tethys' and consequently no 'Tethyan evolution' could have existed before the Pangaea largely formed during the middle Carboniferous. Therefore the history of dispersal of many of the primary components of the Tethyside collage lies outside the Tethyside history *sensu stricto*, because their separation from Gondwana Land predated the middle Carboniferous. If we remove this temporal limitation for a moment, we may view the entire evolution of the Tethyside collage as a diminution in the size of Gondwana Land by continuous calving of continental blocks and their northerly migration to unite with Laurasia.

Fig. 19a shows that, during the late Permian, the terminal Uralide collision was about to occur, and thus the Permo-Triassic Pangaea was in the process of being born. As yet, the entire Cimmerian continent constituted a part of the northern margin of Gondwana Land, with the exception of the eastern part of the Qangtang block (*s''*). Here we place that fragment next to the Yangtze block (*w*), because not only their late Permian biogeographies are similar, but also both contain rift volcanism of basaltic (alkali to transitional types) and silicic types (see Chang *et al.* 1986 for Qangtang, and Fan 1978 for the Yangtze block), known as the Emei Shan basalts in the Yangtze block (Fig. 19a). Fan (1978) interpreted the Emei Shan basalts as products of a marginal-basin opening event. We follow this interpretation and correlate the Emei Shan basalts with the Permian basalts and felsic volcanics of the eastern Qangtang, and interpret them as products of the late Palaeozoic subduction along the future sutures *XXI* and *XXIII*.

This subduction zone may have joined southwards with the one active in the future Iran and Turkey or with the one in the Central Pamirs. The particulars of this possible geometry are shown, for a slightly later time, in the more detailed reconstruction in Fig. 20a, showing the early Triassic palaeogeography. In this reconstruction, the Sanandaj Sirjan zone is interpreted as a north-east-facing magmatic arc that had been active between the late Devonian and the late Permian as a continental-margin

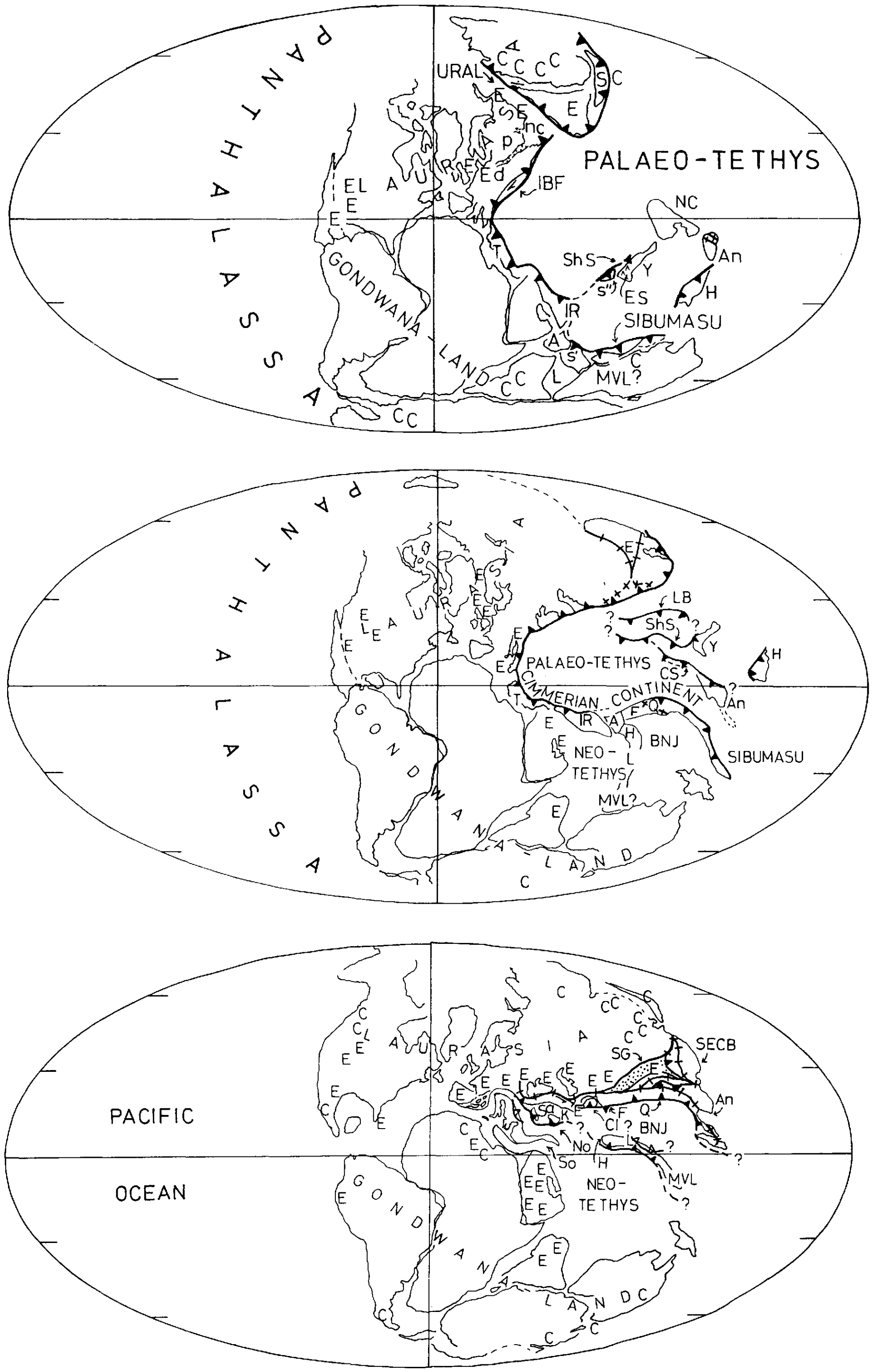
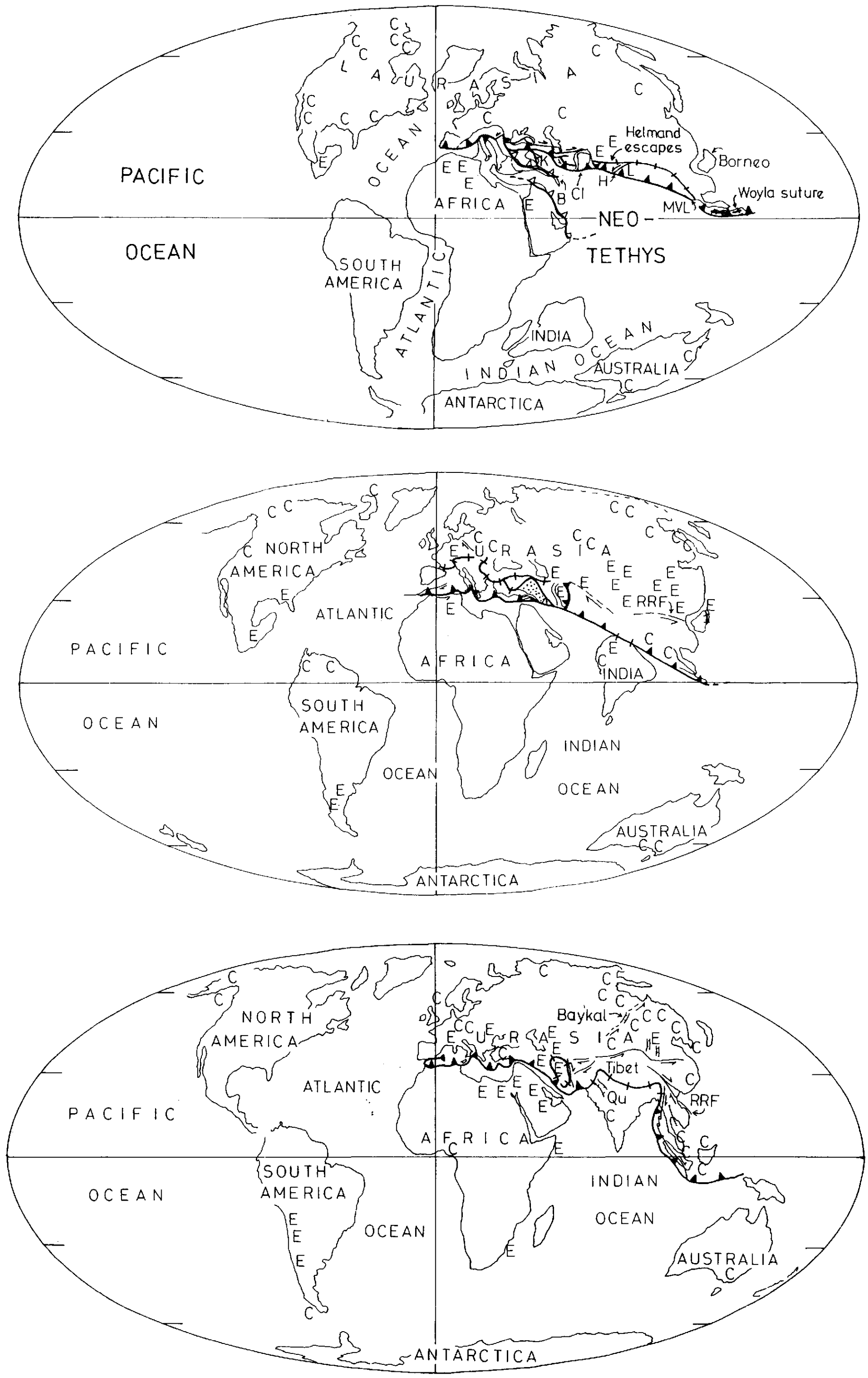


FIG. 19. Highly schematized reconstructions, displaying our present interpretation of the evolution of the Tethyan domain. Abbreviations for all reconstructions in this figure: *A*—Afghan blocks, *An*—Annamia, *B*—Bitlis–Pötürge fragment, *BNJ*—Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean, *CI*—Central Iranian microcontinent, *CS*—Chola Shan, *d*—Dnyepr–Donetz aulacogen, *ES*—Emei Shan, *F*—Farah block, *H*—Helmand block (*sensu* Şengör 1984), *Hu*—Huanan block



s.l., *IBF*—Istanbul–Balkan fragment, *IR*—Iranian block (undisrupted: i.e. Sanandaj–Sirjan zone + north-west Iran and central Iranian microcontinent: see Fig. 20a for details), *K*—Kirşehir block, *L*—Lhasa block, *LB*—Luochou Arc, *MVL*—Mount Victoria Land, *No*—northern branch of Neo-Tethys, *p*—Pachelma aulacogen, *Qu*—Quetta-Sibi graben, *RRF*—Red River fault, *S*—Tarim, *S′*—Pamir–west Qangtang block, *S″*—east Qangtang block, *Sa*—Sakarya Continent, *SG*—Songpan–

continued over

arc, perched on the northern margin of Gondwanaland. Its western continuation is believed to be represented by the pre-Jurassic basement of the Dzirula, Adzharia–Trialeti and the Artvin–Karabagh zones of the Trans-Caucasus and the Eastern Pontides, and the Sakarya Continent of northern Turkey. Within the Adzharia–Trialeti/Artvin–Karabagh zones, the pre-Jurassic basement crops out mainly in the Khrami and the Loku Massifs. This continental-margin arc is here believed to have been responsible also for the late Palaeozoic compressional deformation in Oman, reported by Michard (1982, 1983).

How far this subduction zone extended eastwards is difficult to decide, but the increased unrest in the Djulfa and Svanetia regions, the late Palaeozoic magmatism in the Gondbad-ı Qabus area of the Alborz Ranges (Stämpfli 1978), in the Anarek region of present-day Central Iran (Roman'ko and Morozov 1983), and in the central and south-east Pamirs (Shvolman 1977) suggests to us that it probably was active at least as far east as the Pamirs. Farther east in north-west Tibet data are sparse, and the question remains open.

Farther west, the same subduction zone probably delimited a southerly-moving Istanbul–Balkan fragment along the north-west margin of Palaeo-Tethys (IBF in Fig. 19a). This fragment now consists of the Istanbul Nappe in north-west Turkey and equivalent late Palaeozoic units in southern Bulgaria (see Şengör *et al.* 1984*a*) and contains a portion of a probably north-east-facing Hercynian foreland fold–thrust belt, plus a molasse basin with a late Carboniferous Euramerian flora (e.g. Kerey 1984). Şengör (1987*b*) postulated that this fragment of the Hercynian orogen had originally belonged to the southern margin of Eurasia and constituted the easternmost prolongation of the southern flank of the European Hercynides (Bard *et al.* 1980). It was later detached along a right-lateral strike-slip fault, and moved southwards into its present place in north-west Turkey. This southerly migration of the Istanbul–Balkan fragment must have occurred between the late Carboniferous and the Lias, although the actual time of transport may have been shorter.

In the extreme east of the Tethyan domain, subduction had already started during the Tournaisian, along a south-dipping subduction zone north of Sibumasu (Fig. 19a), and was also probably active during the Permian.

Throughout the Tethysides, normal faulting and rifting-related volcanicity began during the Permian (from mainland Greece, via Crete eastwards: Fig. 4) and may have already achieved considerable separation between Sibumasu and north-west Australia. South of Sibumasu, this extension generated a marine embayment between it and Australia, in which the glacio-marine pebbly mudstones were deposited.

In China, Huanan-ward dipping subduction and associated late Palaeozoic magmatism on the Huanan block were active.

Along the north, Asiatic, margin of Palaeo-Tethys, the blocks *r*, *q*, *p*, *o*, and *n* had already collided with the Altaid nucleus of Asia before the Permian and Asia-ward dipping subduction zones were active along the sutures *VII*, *VIII*, *XIX*, *XX*, *XXXIIII*, and also probably along *XXXIII'* and *XXXV*, although no late Palaeozoic igneous activity has been reported yet from the last two.

In the later early Triassic, the general geometry of the Tethyan domain may have looked something like that shown in Fig. 19b. Notice here that both the Banggong Co–Nu Jiang and the Neo-Tethyan oceans are in the process of active opening, although west of Afghanistan, this opening may still have been in the form of intracontinental stretching, as indicated by the conditions in Oman. In Fig. 19b we have assumed that the Istanbul–Balkan fragment had already reached its final destination in

Ganzi system, *ShS*—Shaluli Shan Arc, *Sibumasu*-China–Burma–Malaya–Sumatra portion of the Cimmerian continent (*sensu* Metcalfe, this volume), *So*—Southern branch of Neo-Tethys, *T*—Turkish blocks.

(a) Late Permian palaeotectonics of the Tethyan domain. Base map is from Scotese (1984). Distribution of coals (*C*) and evaporites (*E*) was taken from Ziegler *et al.* (1979). Map was modified according to data summarized in this paper.

(b) Early Triassic palaeotectonics of the Tethyan domain. Base map and the distribution of coals (*C*), and evaporites (*E*), for this and all subsequent reconstructions, were taken from Parrish *et al.* (1982). This and the subsequent maps were modified according to data summarized in this paper.

(c) Late Jurassic palaeotectonics of the Tethyan domain.

(d) Late Cretaceous palaeotectonics of the Tethyan domain.

(e) Middle Eocene palaeotectonics of the Tethyan domain.

(f) Late Miocene palaeotectonics of the Tethyan domain.

north-west Turkey, but this may be somewhat too early, according to some very preliminary palaeomagnetic data (Sarıbudak *et al.* 1986). The main Song Da subduction zone was active at this time, and may have been gradually propagating westwards to eventually join the active Jinsha Jiang/Nan-Uttaradit/Bentong–Raub subduction zone. North of the Litang subduction zone a new, intra-oceanic arc system, now represented by the Luochou 'arc–trench belt' of Zhang (*XXIII'*: verbal comm. 1985) may have come into being at this time.

Notice in Fig. 19b that we have left the Mount Victoria Land block of Mitchell (in press), still attached to north-west Australia. This fragment may have rifted from this position as late as the late Jurassic (see Şengör 1987; for an alternative and equally viable reconstruction see Hamilton 1983, fig. 1).

Returning to the details of the Iranian case, by the late Triassic, we see that the whole of Iran north-east of the Zagros suture had already been sutured to Laurasia. This suturing event extended between the Greater Caucasus and the Paropamisus, and may have had the geometry shown in Fig. 20b. During this collision, the Farah block probably became severed from (Central) Iran along a strike-slip fault and continued to approach the Paropamisus/Hindu Kush subduction zone.

In the late Jurassic, a considerable portion of the Cimmeride collage had already been assembled (Fig. 19c), and the collisions responsible for this assembly had already disrupted large areas of Eurasia north of the Tethysides. A major segment of the Cimmerian continent north of the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean had collided with Laurasia all the way from northern Greece to the Malayan Peninsula. Although the assembly of the extra-Tibetan Cimmeride collage in China had long been complete, its construction mainly by the activity of the peripheral hidden subduction zones on both sides of the Songpan–Ganzi system continued. Palaeomagnetic evidence shows that the tightening of the suture zones in China and Indochina, during the Jurassic, resulted in a 1650 ± 750 km northward motion of Annamia and South China with respect to 'stable' Eurasia (Achache 1984). An inspection of Figs 5 and 6 clearly shows that much of this tightening was also accomplished by intracontinental shortening and crustal thickening. Notice how narrow most magmatic-arc systems still were during the Triassic and the early Jurassic interval, i.e. before, and just after, most of the Cimmeride collisions (Fig. 5). By contrast, magmatic activity is seen to have spread over much wider areas during middle Jurassic–early Cretaceous times, towards the end of a long period of post-collisional tightening of most of the Cimmeride sutures in China and in the Caucasus (Fig. 6).

Jurassic post-collisional intracontinental convergence is also attested to by the evolution of Jurassic compressional structures and persistent topographic highs, along a belt stretching from eastern Iran to China and following the northern margin of the present-day Alpine–Himalayan system. In fact, Hallam (1984) recently argued that the 'anomalous' late Jurassic arid belt of western and central Asia may have developed in the rain shadow of the Cimmeride mountain wall that rimmed the southern periphery of Eurasia at the time, when monsoonal winds possibly had a more dominant role than today (Şengör 1985*b*).

In the late Jurassic the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean had already begun to close, both by the obduction of a vast ophiolite sheet on to the Lhasa block (*c.* 170 Ma ago), and by subduction. The Mount Victoria Land end of the Helmand/south-west and south-east Pamir/Lhasa block became finally detached from north-west Australia in late Jurassic times.

In the Mediterranean region, the plate boundary systems of the Neo-Tethys and the Atlantic Ocean had linked up already in the early Jurassic, and ocean-floor spreading began in the middle Jurassic, as testified by numerous ophiolite spreading ages in the Apennines, the Alps, and the Carpathians (Fig. 6). Intra-oceanic subduction in the Vardar ocean had also begun by the end of the Lias (Roddick *et al.* 1979), and the Eo-Hellenic nappes were expelled on to the eastern margin of the African promontory in Yugoslavia and Greece in the late Jurassic, possibly across former fracture zones (Şengör *et al.* 1984*a*). Consumption of the floor of Neo-Tethys was also under way in the late Jurassic, from northern Greece, possibly through Turkey and Iran, to the Himalaya.

In the Middle East, the Sanandaj–Sirjan zone and its equivalents in the Trans-Caucasus and northern Turkey had begun slipping left-laterally. This continental-margin-parallel motion had been going on since the early Jurassic, as the opening of the Slate-Diabase mini-ocean in the Greater Caucasus testifies (Beridze 1984). The Sanandaj–Sirjan zone strike-slip system had been bent into a compressional geometry in Iran, and began tearing Central Iran into pieces (Fig. 20c). This led to the opening of the Sevan–Akera–Qaradagh suture in Transcaucasia and northern Iran (Khain 1975, Ber-

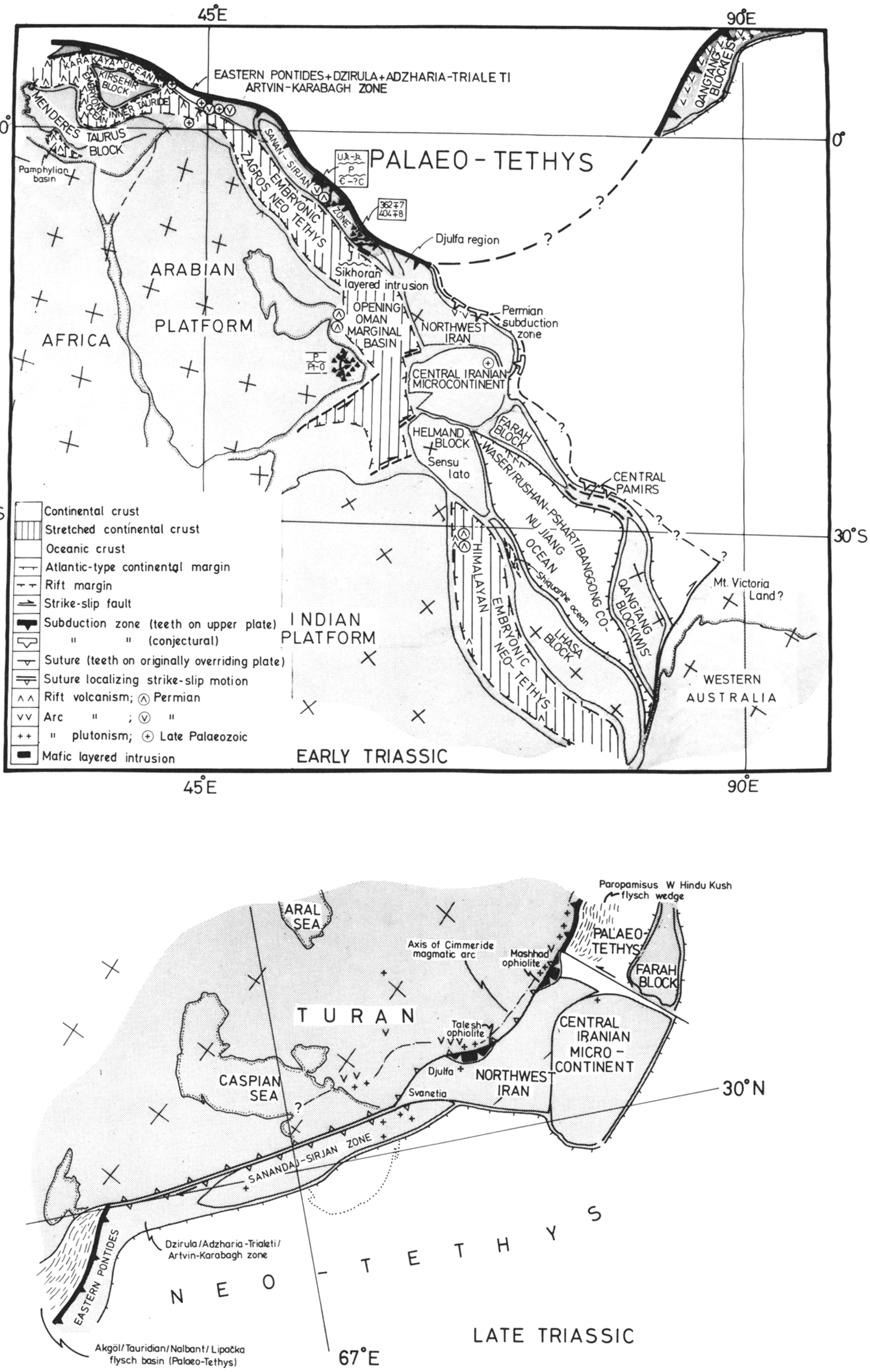
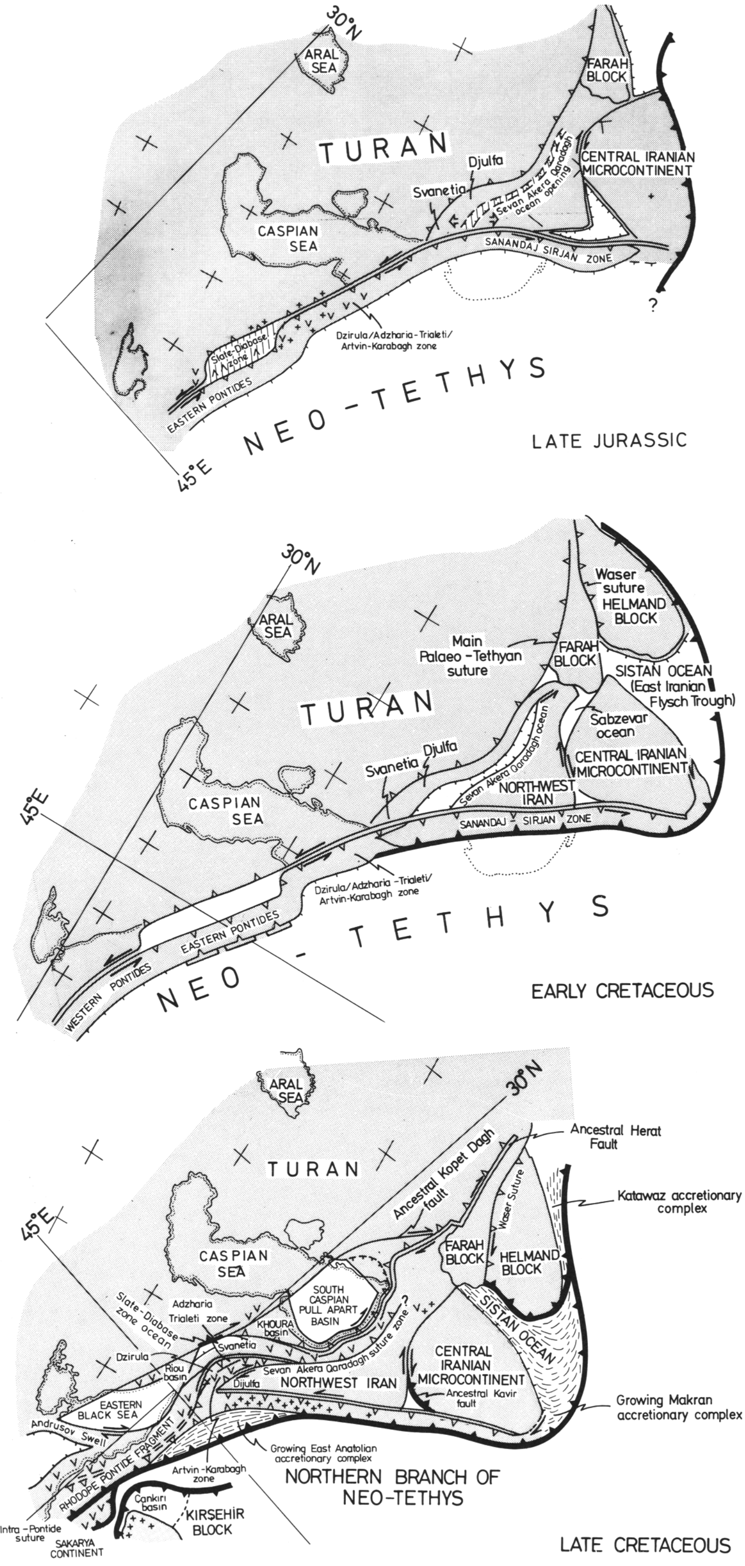


FIG. 20. Tectonic evolution of Iran and surrounding regions, illustrating the complexities introduced by major episodes of strike-slip faulting in the development of a segment of the Tethysides. Comparable complications also exist in other segments (see Fig. 22), where the documentation of the geological evolution is less easy.

(a) Early Triassic (±Griesbachian) palaeotectonics. For sources of data, see caption to Fig. 8a.

(b) Late Triassic (Rhaetian) palaeotectonics of Iran and surrounding regions. For symbols see Fig. 20a.



(c) Late Jurassic palaeotectonics of Iran and surrounding regions. For symbols see Fig. 20a.
(d) Early Cretaceous palaeotectonics of Iran and surrounding regions. For symbols see Fig. 20a.
(e) Late Cretaceous palaeotectonics of Iran and surrounding regions. For symbols see Fig. 20a.

berian *et al.* 1981) and the western part of the circum-Central Iranian Microcontinent ocean. This last event was accompanied by the beginning of rotation of the Central Iranian Microcontinent. Therefore, in this geometry, both the Sevan–Akera–Qaratagh and the western half of the circum-Central Iranian Microcontinent oceans seem to be unrelated to the opening of Neo-Tethys. They are independent, small oceans that formed as a result of large-scale strike-slip tectonics along the southern margin of Laurasia.

Fig. 20d shows a further step in the Mesozoic palaeotectonic evolution of Iran and surrounding regions. Notice here that continuing strike-slip motion between the Sanandaj–Sirjan zone and the rest of Iran further rotated the Central Iranian Microcontinent. One result of this ongoing rotation was the opening of the Sabzevar mini-ocean, as evidenced from early Cretaceous ages obtained from the oceanic sediments in this zone (Lindenberg *et al.* 1984).

In the late Cretaceous, the Lhasa block had been welded to Eurasia already, and the northward consumption of the floor of Neo-Tethys was under way all along the southern margin of Eurasia, from Spain to Indonesia, with the singular exception of the northern margin of the African Promontory, where subduction was southward under the Apulian margin (Fig. 19d). Other north-dipping subduction zones also formed within the Neo-Tethys in Turkey, Iran, and Oman, but they were short-lived, because they originated too close to the north-facing passive continental margins. They collided with these margins and resulted in widespread ophiolite obduction from western Turkey to Oman, along a number of fronts, namely along the sutures *14*, *17*, *18*, *19*, *20*, *21*, *23*, *25*, and *27*. These obduction events mostly took place between the Turonian and Maastrichtian, but at least one example is known from the Kırşehir Massif (*K* in Fig. 19; *h* in Fig. 3), where southerly-vergent (present geographical orientation) ophiolite obduction occurred in pre-Cenomanian times.

The formation of these tremendously long Neo-Tethyan subduction zones, which mostly had originated in the early to middle Cretaceous interval was probably one important cause of the late Cretaceous acceleration of global plate motion rates and the consequent late Cretaceous transgression (cf. Hays and Pitman 1973; Pitman 1978). Widespread ophiolite obduction and a few continental collisions which took place in the late Cretaceous resulted in a loss of at least one-third of the Cretaceous subduction zones in the Neo-Tethyan domain. This may have slowed down the global plate mechanism and have caused the latest Cretaceous–early Cainozoic regression.

The generally compressive tectonics of the late Cretaceous in the entire Tethyan belt, also made itself felt in Iran and most of the small, intra-Laurasian oceanic basins such as the Slate-Diabase Zone in the Greater Caucasus, the Sevan–Akera–Qaradagh ocean in northern Iran, the Sabzevar and related oceans all began closing (Fig. 20e). Possibly as a result of this compressive interference, a sliver of the Sanandaj–Sirjan zone became detached and was inserted sideways into the western end of the Sevan–Akera–Qaradagh oceanic space. This sliver now forms the Adzharia–Trialeti–Artvin–Karabagh (*ATAK*) zone of Transcaucasia and separates the Svanetia sliver of 'northwest Iran' from Djulfa (Fig. 20e). The Dzirula Massif is a small fragment of the *ATAK* zone, possibly somewhat rotated during the insertion of the *ATAK* into its present location. The Chorchana–Utslevi zone in Dzirula, and the Chochiani River ophiolites in the Khrami, are remnants of the Palaeo-Tethyan suture in Transcaucasia disrupted by this episode of strike-slip faulting.

The Sistan ocean also began closing at this time, probably as a consequence of the sideways motion of the Helmand block along the former Waser suture. Another possible consequence of the increased rate of convergence in the Tethysides, during the late Cretaceous, was probably the major right-lateral strike-slip system; the southern Caspian and the eastern Black Sea depressions were opened along its pull-apart segments, or at least their opening was helped along. This pull-apart tectonism also disrupted the former Cimmeride Arc in Turan, and redistributed its pieces over such now widely separated regions as Krasnovodsk, the Gondbad-ı Qabus area in the eastern Alborz Mountains, and Aghdarband.

Fig. 19e shows the middle Eocene (Lutetian) geometry of the Tethyan domain. Notice that along a considerable stretch between the Pyrenees and eastern Turkey, Neo-Tethys and the Alpine branch of the Atlantic Ocean sutured at this time, and the north-west African, Calabrian, Hellenic, and Cyprus subduction zones—connected with the main Neo-Tethyan subduction zone farther east via Zagros—overtook the main convergence between Europe and Africa. In the east, India's north-west corner touched Asia, and the subcontinent began rotating counterclockwise around this point of pinning.

By late Miocene (Vindobonian) time, the Tethysides began acquiring their present-day appearance (Fig. 19f). At about this time both the western Mediterranean (Fig. 19f) and the peri-Arabian Alpide oceans vanished; very shortly thereafter, Timor (*T* in Fig. 19f) began climbing on to the Sahul shelf. Only in the eastern Mediterranean, the Arabian Sea and the Somali Basin, the Bay of Bengal, and in the immediate foreland of the Sumatran subduction systems, do the last remnants of Neo-Tethys still survive and are being actively subducted. Soon, no remnant of the maternal Tethys will be left. Wiens (1985/86) has shown that south of the Indian subcontinent, between the Chagos Bank and the Ninetyeast Ridge, the Indian Ocean is already feeling the birth-pangs of yet another mountain range, which, however, will be no longer of Tethyan descendence and may initiate a new mountain dynasty.

Discussion and conclusions

In the preceding sections we reviewed the available evidence to establish the timing of ocean closing and opening along the sutures of the Tethyside superorogenic complex, and the original provenance and subsequent path of migration of primary components of its vast orogenic collage—forming about one-third of the bulk of Eurasia. The Palaeo-Tethys was born when the Permo-Triassic Pangaea was assembled and therefore it is meaningless to speak of the 'opening of Palaeo-Tethys'. It was born as an embayment, and was essentially a giant gulf of Panthalassa.

We have noticed that not only do the Cimmeride sutures not continue into those of the Hercynides and the Altaids, except in the case of two sutures (southern Ghissar and Mongolo-Okhotsk), but also the Alpides, are entirely independent of the Cimmerides, with the singular exception of the continuity between the Waser and the Sistan sutures. Similarly, when a new mountain range arises from the future obliteration of the Indian Ocean, it will have direct connection with the Alpides along only four sutures out of thirty-five—namely the Zagros, Waziristan, Indo-Burman, and Timor sutures. This simple observation supports Şengör's (1979*a*, 1984) classification of the Tethysides as a distinct orogenic system south of the Hercynides and the Altaids, and his recognition within them of two independent orogenic systems, the Cimmerides and the Alpides.

This recognition has also received independent support (e.g. Belov *et al.* 1985, 1986; Chang *et al.* in press). In the tectonic evolution of the Tethysides, we saw that the elimination of Palaeo-Tethys formed the huge, multi-branched orogenic zone of the Cimmerides, as a consequence of the collision with Eurasia, not only of the Cimmerian continent, but also of an immense orogenic collage that had accreted mostly around its eastern end. By contrast, the closure of Neo-Tethys involved a much smaller number of continental blocks, and the resulting Alpides consequently include a much smaller orogenic collage. In fact, much of the Alpide deformation actually took place on a Cimmeride foundation. This overprinting is the main reason why the recognition of the Cimmerides as an independent Tethyan orogenic system has been so difficult.

We have seen further that Belov's (1981) suggestion of treating the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay suture as a 'separate' Meso-Tethys (in the sense of a 'middle Tethys'), independent of both Palaeo- and Neo-Tethys, cannot be supported, despite this ocean's considerable size, because it did not extend any farther west than eastern Iran, but also, in Tibet, it was directly connected with Palaeo-Tethys via the mid-Qangtang suture and actually formed a part of it. Correlations by Belov (1981), of suture segments farther west than eastern Iran, with his Meso-Tethys (e.g. in the Caucasus) similarly do not agree with the available observations, as the previous discussions show.

We found subduction-related magmatism to be a generally reliable guide for the detection and tracing of suture zones. Where the clear linear/arcuate magmatic arcs seem to be interrupted for reasons other than just younger cover—as in the most conspicuous example of the late Palaeozoic to Triassic Turco-Iranian magmatic gap or the late Palaeozoic West Thailand magmatic gap (Figs 4 and 5), we realized that it was commonly a result of post-collisional strike-slip-related disruption of a formerly more regular pattern (Figs 20a–e).

There are also primary magmatic gaps in our maps, such as the North China magmatic gap of Triassic and early Jurassic age (Fig. 5). Such primary gaps in the continuity of subduction- *and* collision-related magmatism, if major (i.e. more than 1000 km), indicate original absence of sutures. In the case of the North China magmatic gap, the interruption corresponds with continental regions separating the Altaid system from the Tethysides.

The character of the *distribution* of convergent magmatism is also diagnostic of tectonic phenomena. Subduction-controlled magmatism generally follows a well-defined, continuous, and narrow band, never wider than a few hundred kilometres, and only occurs on one side of the associated suture. Best examples of this are seen in our maps showing late Palaeozoic and late Cretaceous–early Tertiary magmatic activity, i.e. at times when a minimum of continental collisions occurred in the Tethyside system. By contrast, times of major collisions and intracontinental convergence are also times of generally patchy magmatism occurring in short but wide bands, or even in equant areas commonly extending on both sides of sutures. Examples of such cases are abundantly present in Fig. 6, mainly in East Asia, but also in the Caucasus. In a few cases, subduction-related magmatism, originally confined to only one side of a suture, may appear to occur later on both sides, following pervasive conjugate wrench faulting during continuing suture-perpendicular shortening (Fig. 21). Examples of such cases are present in the Kuen-Lun and in the Qin-Ling Range, and also in eastern Burma and West Thailand.

In sharp contrast to convergent-boundary magmatism, rift-related magmatism appears to be a poor guide for delineating former axes of rifting. If rift magmatism was the only guide, we would have hardly recognized the continuity—in places, even the presence—of the Neo-Tethyan rifted margins. This is hardly surprising when compared with regions of active rifting. Sparsity of rift magmatism is also seen, both along the margins of the Atlantic Ocean (Burke 1976) and along the East African Rift Valleys (Burke and Whiteman 1973).

A further difficulty in the recognition of 'opening'-related magmatism in the Tethysides was encountered in the central and eastern Mediterranean region. As seen in Fig. 4, abundant mafic volcanism characterized much of post-Hercynian Europe and was apparently related to post-collisional phenomena, and/or to a broad shear movement that affected much of Europe in post-Hercynian times (Arthaud and Matte 1977). There seems, however, to have been no natural spatial break between this 'convergent' basaltic volcanism, and the 'rift-related' volcanism in mainland Greece and

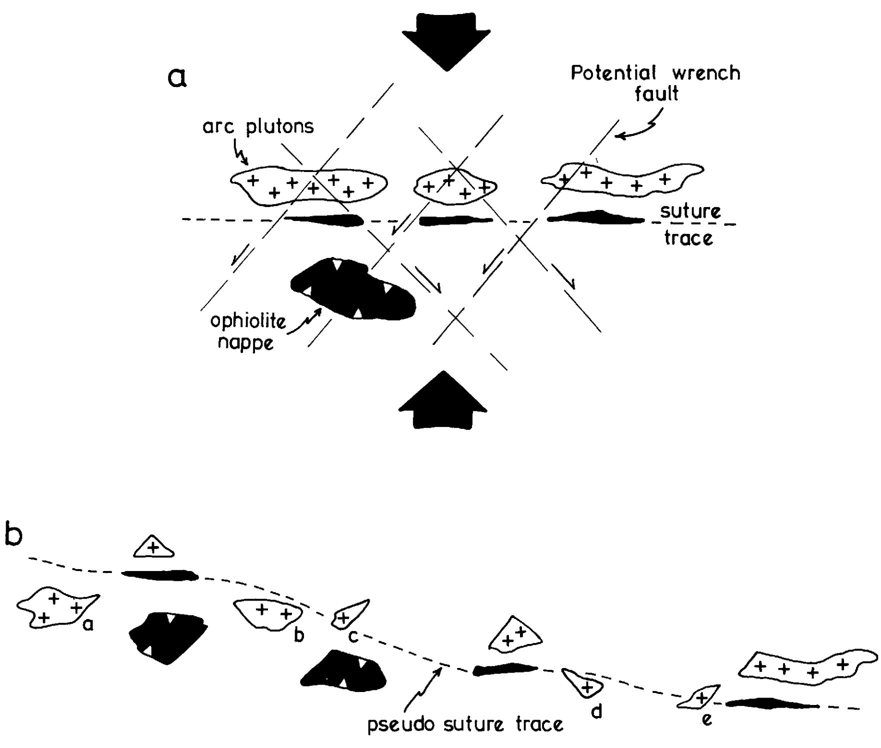


FIG. 21. A mechanism to generate 'pseudo-suture-crossing granites' as seen in northern Tibet. This mechanism also serves to disrupt, and make difficult, the recognition of old sutures.

(a) Before conjugate wrench-fault-dominated post-collisional shortening.

(b) After conjugate wrench-fault-dominated post-collisional shortening. Here the arc-related plutonic rocks *a*, *b*, *c*, *d*, and *e* appear as if they were post-collisional, suture-crossing granitic rocks.

Crete. It is extremely difficult to decide now whether the Cretan late Palaeozoic basaltic volcanism was somehow related to Hercynian events, or to Cimmeride 'back-arc', i.e. rifting events. We prefer the latter interpretation, because the Hercynian orogenic belt did not extend to Crete.

An inspection of Figs 4 through 7, and 19a through 19f, shows how collisions seem to have triggered the generation of new subduction zones. The early to late Carboniferous Altaid collisions in Central Asia probably triggered the coeval onset of the destruction of Palaeo-Tethys; the collision of the Farah–Central Pamir–Western Qangtang fragment (*l,s*) with Laurasia in the early Jurassic seems to have led to the beginning of consumption of the Waser/Rushan–Pshart/Banggong Co–Nu Jiang ocean, probably in the middle Jurassic. The early Cretaceous elimination of the latter greatly accelerated the subduction of Neo-Tethys that had probably begun earlier as a result of ophiolite obduction on to the Lhasa block. Similarly, the Eocene elimination of Neo-Tethys along the Himalayan segment, and the continuing intracontinental convergence since then, appear to be causing the inception of yet another consuming margin south of India at the present day: very much like the earlier cases since the late Palaeozoic. In other segments of the Tethysides, similar relationships between collisions and inception of subduction zones are also seen.

Not only subduction zones may be initiated by collisions, but also spreading centres may be similarly triggered by subduction zones. McKenzie and Weiss (1975) suggested that major spreading centres may be born initially in back-arc basins behind subduction zones, and Şengör (1979) argued that in some places Neo-Tethys may have opened as a series of back-arc basins above Palaeo-Tethyan subduction zones. Later, Şengör *et al.* (1980) have shown that, at least in northern Turkey, the northern branch of Neo-Tethys opened by splitting the Palaeo-Tethyan continental-margin arc. The data reviewed previously now suggest that a large proportion of the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay ocean and the Emei Shan marginal basin may also have opened as back-arc basins above south-dipping subduction zones in the Central Pamirs, easternmost Qangtang, and western Thailand. Not only in Turkey, but also in Iran and in Oman, Neo-Tethys appears to have begun as a marginal basin behind the south-west-dipping Palaeo-Tethyan subduction zone. As documented above, there is now increasing evidence to suggest that, in fact, much of the Gondwanan north and north-east margin facing Palaeo-Tethys may have been active, and that the initial disintegration of this margin to open both the Waser/Rushan–Pshart/Banggong Co–Nu Jiang/Mandalay and Neo/Tethyan oceans may have been due entirely to back-arc basin activity.

Tethyside sutures stitch together the Tethyside blocks that form the primary components of the Tethyside orogenic collage. In search of clues concerning the original provenance of these blocks, we have looked at their late Carboniferous to early Cretaceous palaeobiogeographical and late Carboniferous to early Permian palaeoclimatological record. The latter record clearly indicates that all of the Cimmerian continent in Asia east of Iran (*i, j, k, l, m, s sensu lato*, except *s″*, *t sensu lato*) with the exception of the eastern part of the Qangtang fragment (*s″*) formed the northern margin of Gondwana Land until at least the middle Permian. In the late Permian the *Neoschwagerina* and *Verbeekina* provinces clearly delineate the Cimmerian continent *and* the Chinese blocks, with the exception of *n, o, p, r*, and some places on the northern shelf of Gondwana Land, suggesting that the whole of Turkey and Iran, including Transcaucasia, belonged to the Cimmerian continent, which itself constituted the northern margin of Gondwana Land, a conclusion fully supported by geological observations (e.g. Şengör *et al.* 1984*a*; Belov *et al.* 1986).

The late Carboniferous to early Permian floral record suggests that the North China (*q*), Yangtze (*w*), Huanan (*y*), eastern Qangtang (*s″*), eastern Qaidam (*o*), Annamia (*x*), and with it the eastern Malayan Peninsula (*x‴*) and south-east Sumatra? (*x⁗*) were all near the equator, a conclusion largely corroborated by palaeomagnetic evidence (Lin *et al.* 1985*a*; McElhinny 1985; Opdyke *et al.* 1986), and had a common Cathaysian flora. This picture remained probably similar in the late Permian, as evidenced by the coincidence between the Cathaysian floral and the *Lepidolina* foraminiferal provinces, as shown in Figs 14 and 15. In the middle Permian, this conclusion is further strengthened by the *Monticulifera–Urushtenoidea* brachiopod subprovince, within what Nakamura *et al.* (1985) call the 'Cathaysia–Tethyan subprovince'. Although the geological record shows that these blocks were terminally sutured only by the late-middle to late Triassic, direct 'isthmian' links between most of them probably existed via island-arc systems, as shown in Fig. 19b, although geological evidence is not now sufficient to elaborate this hypothesis. Separation until the Triassic, between the North China and Yangtze blocks is suggested, in addition to palaeomagnetic (Lin *et al.* 1985*a*: McElhinny 1985, Opdyke *et al.* 1986) and geological (Şengör 1985*c*) evidence, by the sharp boundary between the northern and southern subprovinces of the Cathaysian floral realm along the Qin-Ling/ Dabie Shan suture zone. In this connection, the northern Cathaysia-type flora of the Qinghai Nan Shan (Fig. 14) is interesting, and may suggest a strike-slip emplacement of this fragment westward into its present location west of the Alxa block from a position farther east, perhaps originally north of the present Qin-Ling, in post-Permian times. A similar strike-slip emplacement of the Istanbul–Balkan fragment in north-west Turkey may have brought the westernmost Turkish occurrence of a late Carboniferous Euramerian flora to its present 'Gondwanan' position (i.e. south of the main Palaeo-Tethyan suture), although similar floras in north-west Turkey, and in Transcaucasia, cannot be explained by the same mechanism. Another possible left-lateral strike-slip displacement is also suggested by the S-shaped *Neoschwagerina* province boundary in western Turkey and Cyprus (Fig. 15).

Such 'anomalous' palaeobiogeographical observations, and also consideration of facies, structure, and even palaeomagnetic evidence suggest that major strike-slip motion in fact did take place along most of the Tethyside sutures before, during, and after collision. Fig. 22 and Table 1 catalogue the best known examples of these. It seems that over 60 per cent (by length) of the Tethyside sutures had considerable strike-slip motion on them. More than 25 per cent had displacements exceeding 1000 km, and over 50 per cent had more than a few hundred kilometres. At least 15 per cent of them experienced more than one major episode of sideways motion. Table 1 shows that, within the

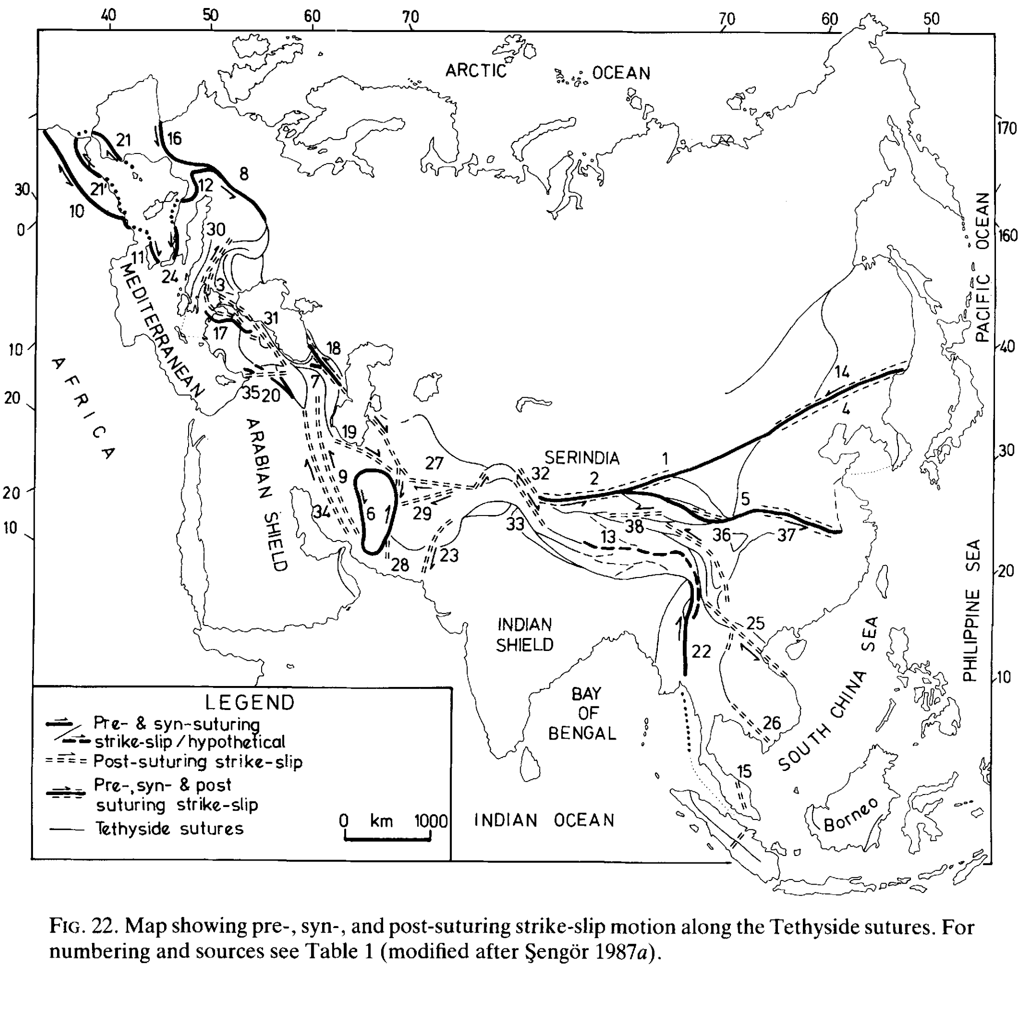


FIG. 22. Map showing pre-, syn-, and post-suturing strike-slip motion along the Tethyside sutures. For numbering and sources see Table 1 (modified after Şengör 1987*a*).

TABLE 1. *Sense, amount, and timing of strike-slip motion on some Tethyside sutures**

Suture†	Sense	Amount	Time	Reference
1 (*XIII*)	Left	?	ePale	Yang *et al.* (1986)
2 (*XI+XIII*)	Left	?	?C–P	Bally *et al.* (1980)
3 (*II*?)	Right	>1000 km	C–?eTr	Şengör (in press, *a*)
4 (*XVI*)	Left	>4000 km	P	Lin *et al.* (1985*a*)
5 (*XX–XXXIII*)	Left	4300 ± 1200 km	P–eJ	Lin *et al.* (1985*a*)
6 (*26*)	Right	~1000 km	lTr–lK	Davoudzadeh and Schmidt (1984)
7 (*VI*)	?	?	P–lTr	Adamia and Belov (1984)
8 (58)	Left	~1500 km	lTr–eK	Weissert and Bernoulli (1985)
9 (*IV'*)	Left	~2000 km	lTr–eK	This chapter
10 (*4–5*)	Left Right	10–20 km 10–20 km	e–mJ lJ–eK	Mattauer *et al.* (1977)
11 (*7* south)	?	?	eJ	Catalano and D'Argenio (1982)
12 (*7* north)	?	?	m–lJ	Gianelli and Principi (1977)
13 (*XXIV*)	Left	>500 km	lJ–eK	This chapter
14 (*XIV*)	Left	?	lJ–eK	Klimetz (1983)
15 (*XXXI*)	?	?	Tr–K	Şengör (1986*b*)
16 (*1*)	Left	150–200 km	e–lK	Trümpy (1976)
17 (*14*)	?	?	lK	Şengör *et al.* (1984*a*)
18 (*24*)	Right Left	~2000 km 350–700 km	lTr–lK lK–Pal	This chapter Şengör *et al.* (1984*b*)
19 (*VI* + *VII*)	Right	~500 km	lK–Pal	Şengör *et al.* (1984*b*) and this chapter
20 (*18,22*)	Right	~1000 km	lK–eE	Y. Yılmaz and O. Sungurlu (pers. comm., 1984)
21 (*2*)	Right	~300 km	lE–lM	Wildi (1983)
21' (*3*)	Left			
22 (*XL*)	Right	~460 × km	lOl–present	Curray *et al.* (1982)
23 (*28*)	Left	>200 km	e–mT	Lawrence and Yeats (1979)
24 (*7* south)	(north) Left (south) Right	~50 km	T	B. D'Argenio (pers. comm., 1980)
25 (*XXIX*)	Left Right	~500 km A few tens of km	O–M Q	Tapponnier *et al.* (1986) Allen *et al.* (1984)
26 (*XXVII*)	Left	~300 km	O–M	Tapponnier *et al.* (1986)
27 (*VIII*)	Right	?	end E–?M	Tapponnier *et al.* (1981)
28 (*26*)	Right	?	post eM	Tirrul *et al.* (1983)
29 (*IX*)	Right	?	end E	Tapponnier *et al.* (1981)
30 (*12*)	Right	>100 km	M–Pli	Burchfiel (1980)
31 (*13,17*)	Right	>100 km	lM–present	Şengör (1979*b*), A. M. C. Şengör and N. Görür (unpublished data)
32 (*XI,XIII*)	Left	500–600 km	M–present	Tapponnier *et al.* (1981)
33 (*31*)	Right	~200 km	lT	Molnar and Tapponnier (1978)
34 (*25*)	Right	60 km	Pli–present	Tchalenko and Braud (1974)
35 (*21,22*)	Left	~20 km	Pli–present	Şengör *et al.* (1985)
36 (*XXIIII*)	Left	?	lT	Molnar and Tapponnier (1975)
37 (*XXXI*)	Left	several hundred km	post–E	Peltzer *et al.* (1985)
38 (*XIX*)	Left	?	lT	Molnar and Tapponnier (1975)

*Abbreviations: Pale—Palaeozoic, C—Carboniferous, P—Permian, Tr—Triassic, J—Jurassic, K–Cretaceous, T—Tertiary, Pal—Palaeocene, E—Eocene, Ol—Oligocene, M—Miocene, Pli—Pliocene, e—early, m—middle, l—late.

†In the enumeration of sutures, left-hand numerals refer to those in Fig. 23 and right-hand numerals (in parentheses) refer to those in Fig. 4.

Tethysides, strike-slip motion along the sutures took place at all times, with no recognizable temporal preference.

Complications introduced by strike-slip movement are not confined to cases along sutures. Collision-induced escape tectonics (Burke and Şengör 1986) also generates a very complicated strike-slip regime (cf. Şengör *et al.* 1985) on all scales. Major escape-related strike-slip faults that formed or moved during the Tethyside evolution are shown in Fig. 2. Some, but not all, of these are identical with some of the structures depicted in Fig. 22. If one now superimposes Figs 2, 3, and 22, one may begin to appreciate the degree of disruption experienced by not only the original Gondwanan northern margin, which was transported in separate parts northwards and reassembled in a different arrangement to form the Tethyside collage, but also the disruption of the Tethyside collage itself during and after its reassembly.

The imaginary map that the reader is invited to draw in his or her mind's eye, by superimposing Figs 2, 3, and 22 would be a 'suspect terrane map' of the Tethysides. If we now took the further step of considering the innumerable terranes in our imaginary map as having been originally independent entities populating the floor of 'the Tethys', as has been done in other orogenic belts (e.g. Coney *et al.* 1980; Jones *et al.* 1983; Schermer *et al.* 1984; Howell *et al.* 1985), the chaos and confusion this would precipitate would be easy to visualize. That is why in this paper we avoided the recommended 'terrane analysis' and instead began by identifying suture-bound blocks by carefully searching for past evidence of *subduction* and *continental collision*. Suturing by strike-slip apposition of blocks is also possible, as Dewey *et al.* (1986) have shown, but it is not common.

In the present highly disrupted state of the Tethyside collage, one good guide to reconstruction of the pre-dispersal picture of the Tethyside blocks, was seen to be the structure and history of the pre-Tethyside basement. In this regard we found the Pan-African structures to be particularly promising and already helpful in the Middle East, where they are reasonably well-known, both on the Gondwanian foreland (North Africa and Arabian Peninsula), and on the individual Tethyside blocks around it (Şengör *et al.* 1984*b*). Elsewhere in Eurasia, knowledge of the pre-Tethyside Pan-African basement is unfortunately very fragmentary.

Finally, we have seen once more, as frequently emphasized by Şengör (1984, 1985*a*, 1986*a*,*b*) the extremely misleading role of the tripartite subdivision of all Phanerozoic orogenic phenomena into Caledonian, Hercynian, and Alpine 'cycles'. This regrettable terminology of the old fixist model of episodic and synchronous world-wide orogeny unfortunately still pervades much of the local and regional Tethyan geological literature, and leads to serious and widespread misunderstanding and confusion, as documented by Şengör (1985*a*,*b*, 1986*a*,*b*).

We presented six palaeogeographical maps of the entire Tethyan domain, to sketch its evolution between the Permian and the late Miocene. A comparison of these maps with a generalized Mesozoic eustatic sea-level change curve (e.g. see fig. 9 in Şengör 1985*b*) reveals a remarkable correlation of times of major Cimmeride collisions with those of major falls in sea-level. Dewey and Windley (1981) argued that because collision gathers the continental crust into a smaller area, it simultaneously enlarges the capacity of the ocean basins. Moreover, because subduction zones, which facilitate plate motion, are lost during collisions, a general slowing down of global plate motions is observed at times of major collisions. Thus, major regressions coincide temporally with important collision events. It appears that the Cimmeride collisions may have provided one possible factor affecting the early and middle Mesozoic sea-level changes.

Figs 19a–f also disclose a persistent trend in the evolution of the Tethysides since at least the Permian, namely the progressive disintegration of the southern supercontinent, and the northerly flight of its dispersed pieces to unite with Eurasia. This tendency to go north is also seen in the blocks that have been accreted to Korea, Japan (e.g. Lee 1984), and to the North American Cordillera (Howell *et al.* 1985), also since the Permian. This large-scale and persistent migration northwards of continental blocks appears significant, and must somehow be related to the kinematics of the first-order convective circulation, at least in the upper mantle. The history of the Tethysides may thus provide an important constraint for models of convective circulation in the mantle.

Another important point that emerges from our survey of Tethyside geology is the abundance of long and thin strips of continental material, that were successively detached from Gondwana Land during the early stages of the evolution of the Tethysides. Similar long and thin 'ribbon continents' are also abundant in the structure of the Altaid Asia (Şengör 1987*b*). The role of such ribbon continents in the evolution of orogenic collages may be more important than hitherto realized. Vink *et al.* (1984) and

Steckler and ten Brink (1986) suggested a mechanism to show why strip-continents may be preferentially produced during continental rifting.

The Mesozoic evolution of Iran suggests that uncertainties, in the order of 500 to 1500 km, are thus implicit in many places in our Figs 19a–f, with respect to initial reconstructions, motion paths of the collage components, palinspastic restoration of block shapes, and post-collisional distortion of entire collages and the host continents around which they accrete. These are significant dimensions implying, at one extreme, that our present reconstructions may be grossly in error, which could only be avoided if we knew the regional geological details uniformly throughout the Tethysides at more than a reconnaissance level. What is, therefore, most needed in the Tethysides now is more, and more detailed, geological fieldwork to map their suture zones and establish better geological criteria for the place of origin and paths of migration of the blocks they bound. Ages, senses, and amounts of major strike-slip displacement along the Tethyside sutures are among the most outstanding problems in the study of the Tethyside collage. Before all the suture zones, and the major strike-slip boundaries within the Tethysides, are mapped in detail, it will not be possible to interpret satisfactorily the available palaeobiogeographical, palaeoclimatological, and palaeomagnetic data to increase the precision of our reconstructions. A corollary to this statement is that: *all palaeobiogeographical, palaeoclimatological, and palaeomagnetic data must be carefully plotted on suture maps or in reference to sutures*, for them to be useful for work on continental reconstructions.

Most importantly, we hope that this study has underlined sufficiently the necessity of a multidisciplinary, iterative approach to regional tectonic analysis, including regional stratigraphy and structure, petrology, palaeobiogeography, palaeoclimatology, and palaeomagnetism, *in the order given*. This multidisciplinary approach to historical geology, with field data as its fundamental basis, remains as relevant today as when it was first formally introduced into geology by Sir Charles Lyell.

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A. M. C. ŞENGÖR, İTÜ Maden Fakültesi, Jeoloji Bölümü, Teşvikiye, Istanbul, Turkey. Also at TÜBITAK-TBAE, Yerbilimleri Bölümü, Gebze, Kocaeli, Turkey.

ALTAN CIN, TIMUR USTAÖMER, İTÜ Maden Fakültesi, Jeoloji Bölümü, Teşvikiye, Istanbul, Turkey.

DEMIR ALTINER, Ortadoğu Teknik Üniversitesi, Mühendislik Fakültesi, Jeoloji Bölümü, Ankara, Turkey.

K. J. HSÜ, Geologiches Institut der ETH-Zentrum, 8092 Zürich, Switzerland.