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**Paleomagnetic study on the late Paleozoic events of the
southern Central Asian Orogenic Belt and constraints on its
tectonic and paleogeographic reconstructions**

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Résumé étendu en français

La Ceinture Orogénique d'Asie Centrale (Central Asia Orogenic Belt - CAOB) ou encore Chaîne des Altaïdes, se situe entre le Craton de Sibérie au nord, le Craton d'Europe (Baltique) à l'ouest, les cratons du Tarim et de Chine du Nord au sud. Elle s'étend sur ~5500 km de l'Oural à l'ouest à la Mer d'Okhotsk à l'est. C'est le système orogénique d'accrétion/collision Phanérozoïque le plus étendu au monde. Depuis plusieurs décennies la CAOB a suscité beaucoup d'attention de la part de la communauté des géosciences et à la suite des nombreux travaux qui y ont été effectués, il existe un consensus selon lequel la CAOB est un vaste complexe constitué de nombreux microcontinents, monts sous-marins, arcs insulaires et prismes d'accrétion. L'amalgame de ces divers éléments s'est réalisé lors de multiples épisodes d'accrétion et de collision au cours du Néoprotérozoïque et du Paléozoïque. Lors de la fermeture finale de l'Océan Paléo-Asiatique au Paléozoïque supérieur – Mésozoïque inférieur, l'évolution de la CAOB a culminé avec l'accrétion des cratons de Chine du Nord et du Tarim le long de sa bordure méridionale. En fin d'évolution, l'accrétion a été suivie d'une importante déformation intracontinentale par réactivation transcurrente des discontinuités préexistantes. Cette déformation intracontinentale a eu pour résultat des déplacements latéraux pouvant atteindre plusieurs centaines de kilomètres. Pendant le Mésozoïque et le Cénozoïque, la CAOB s'est trouvée dans le champ lointain de collisions successives (Lhassa, Tibet, Inde). La collision Inde-Eurasie amorcée à l'Eocène, est à l'origine du raccourcissement général et de la surrection de la CAOB.

De nombreux travaux récents ont apporté des éléments clés sur l'évolution de la CAOB, y compris le caractère polyphasé de la tectonique d'accrétion/collision du Néoprotérozoïque au Paléozoïque, la transition tectonique pendant le Paléozoïque supérieur – Mésozoïque inférieur, ainsi que la tectonique intracontinentale pendant le Mésozoïque et le Cénozoïque. Cependant, en raison de sa complexité et d'investigations parfois insuffisantes, la compréhension de l'évolution du Paléozoïque

supérieur au Mésozoïque inférieur reste incertaine. C'est dans la partie du Sud de la CAO, incluant la Ceinture Orogénique du Tianshan (Tianshan Orogenic Belt - TOB), la Ceinture Orogénique de Beishan (Beishan Orogenic Belt - BOB), et la Ceinture Orogénique Mongolienne (Mongolia Orogenic Belt - MOB), que se trouvent des preuves de la fermeture finale de l'Océan Paléo-Asiatique et de l'amalgame final avec les cratons du Tarim et de Chine du Nord. Par conséquent, cette région revêt la plus grande importance pour la compréhension des évolutions tectonique et paléogéographique de la CAO.

Les problèmes majeurs dans le Sud de la CAO à résoudre pour atteindre un consensus sont les suivants :

- Identification de la cinématique et mise en évidence de zones décrochantes à grande échelle
- Compréhension de la transition entre tectonique de collision/accrétion et tectonique intracontinentale
- Réalisation de reconstructions paléogéographiques bien contraintes du Paléozoïque supérieur au Mésozoïque inférieur

(1) Identification de la cinématique et mise en évidence de zones décrochantes à grande échelle

D'après les données paléomagnétiques, la magnitude des déplacements décrochants le long de zones de cisaillement majeures dans la TOB atteint plusieurs centaines de kilomètres. Ces zones de cisaillement dextre couplées avec la zone d'Erqis en cisaillement sénestre indiquent une extrusion vers l'Est des unités tectoniques de Yili, Junggar et du Tianshan du Nord par rapport aux cratons de Sibérie et du Tarim, en raison probable d'une convergence oblique intraplaque. Des zones décrochantes de grande échelle sont également développées dans les BOB et MOB, mais elles sont moins étudiées. Des études récentes de paléomagnétisme dans la MOB suggèrent que

des rotations significatives liées aux mouvements décrochants ont eu lieu après le Permien inférieur autour d'axes verticaux situés à l'intérieur et sur les frontières du bloc unifié constitué par les blocs de Mongolie et de Chine du Nord, même si leur cinématique est mal contrainte.

De nombreux chercheurs considèrent ces déplacements décrochants comme l'ajustement intracontinental après la fermeture du Paléo-Océan d'Asie. En effet, les failles décrochantes du Sud de la CAO B forment un système orienté ouest-est avec des âges et cinématiques similaires. Un tel système de failles implique qu'un mouvement décrochant de grande ampleur est le mode principal de la tectonique intracontinentale après l'amalgame final de la CAO B. Ce système joue un rôle important dans la configuration finale de la partie sud de la CAO B, est lié à l'activité magmatique du Paléozoïque supérieur – Mésozoïque inférieur ainsi qu'aux minéralisations associées, reflétant ainsi la géodynamique régionale. De plus, ces zones de cisaillement offrent une excellente opportunité pour l'étude la répartition des contraintes et la localisation des déformations en transpression et en transtension à l'échelle régionale. Il est clair qu'une densification des sites d'étude est nécessaire en vue d'une analyse cinématique de ces zones décrochantes et de leur implication dans la tectonique à grande échelle de la partie sud de la CAO B.

(2) Compréhension de la transition entre tectonique de collision/accrétion et tectonique intracontinentale

La compréhension de la transition de la tectonique de collision/accrétion vers la tectonique intracontinentale reste confuse. Le calendrier, les modalités et les processus géodynamiques en jeu lors de cette transition ont été longtemps l'objet de controverses. Différentes époques pour la fermeture de l'océan Paléo-Asiatique ont été proposées aux seins ou entre des segments du sud de la CAO B. Différentes hypothèses, dont la convergence oblique, le retrait de la plaque plongeante, et l'extrusion tectonique, ont été émises pour interpréter cette transition tectonique pendant la période fini-

Paléozoïque – Mésozoïque supérieur. Par ailleurs, différents contextes géodynamiques ont été suggérés pour le magmatisme du Paléozoïque supérieur, à savoir, marge active, subduction de dorsale, extension/rifting post-orogénique, et volcanisme intraplaque de plume mantellique lié à la LIP (Large Igneous Province). La difficulté d'atteindre un consensus est due probablement à la spécificité des approches appliquées à ces sujets (principalement géochronologie et géochimie du magmatisme), qui conduisent souvent des interprétations tectoniques ambiguës. Il est donc nécessaire d'élargir les points de vue et d'offrir de nouvelles perspectives disciplinaires.

La cinématique des déplacements relatifs le long des bordures des microblocs continentaux dans le sud de la CAO B au paléozoïque supérieur – mésozoïque inférieur est une clé pour le déchiffrement de l'évolution géodynamique de la transition de la subduction/accrétion vers la tectonique intracontinentale. Cependant, ces bordures ont subi une évolution longue et polyphasée ce qui en rend la reconstruction géodynamique compliquée et objet de controverse. Le paléomagnétisme est une méthode efficace pour résoudre ce problème car il fournit des contraintes quantitatives sur les mouvements de blocs, y compris la translation latitudinale et les rotations autour d'axes verticaux. La comparaison des pôles paléomagnétiques des différentes unités litho-tectoniques du sud de la CAO B fournit donc une base essentielle pour la compréhension de cette transition tectonique.

(3) Réalisation de reconstructions paléogéographiques bien contraintes du Paléozoïque supérieur au Mésozoïque inférieur

Il est couramment admis que les positions des principaux cratons constituant la Pangée sont suffisamment bien encadrées par de nombreuses données paléomagnétiques, géologiques et géochronologiques. Cependant, la position des blocs de l'Asie orientale dans la Pangée reste controversée. La première cause en est la complexité des structures géologiques, particulièrement dans la CAO B, composée de nombreux microcontinents, arcs insulaires, monts sous-marins, et prismes d'accrétion,

tous variablement déformés. A cause de cette complexité, et à l'exception du Bloc du Tarim, peu de données paléomagnétiques fiables avaient été obtenues. Par conséquent, aucune reconstruction paléogéographique robuste n'avait été réalisée.

Néanmoins, la paléogéographie est essentielle pour comprendre l'origine et la diversité de la vie, les paléoclimats, et la distribution des ceintures tectoniques et des ressources minérales. Plus spécifiquement, les positions relatives des unités tectoniques au sud de la CAOBS sont la base de la restauration et de la reconstruction de son contexte océan-continent, l'évolution des bassins océaniques et la 'dérive' continentale. Les progrès des recherches récentes dans l'Himalaya présentent une bonne analogie de cet aspect. Il est donc nécessaire de construire une paléogéographie bien contrainte pour la partie du Sud de la CAOBS pour la période du Paléozoïque supérieur – Mésozoïque inférieur.

Des données paléomagnétiques de haute qualité joueront un rôle essentiel pour résoudre ces problèmes c'est pourquoi l'étude paléomagnétique a été couplée à des observations de terrain et diverses méthodes d'investigation. Ainsi, la pétrographie sur lames minces, le magnétisme des roches, l'anisotropie de susceptibilité magnétique (ASM), la microscopie électronique à balayage (MEB), la géochronologie et la thermochronologie ont été appliquées à des dykes basiques d'âge néoproterozoïque et permien, des migmatites carbonifères, des roches volcaniques et sédimentaires permien provenant des blocs de Yili, Turpan-Hami et Baluntai dans la TOB, aussi de la zone de Shuangyingshan dans la BOB (Figure 1).

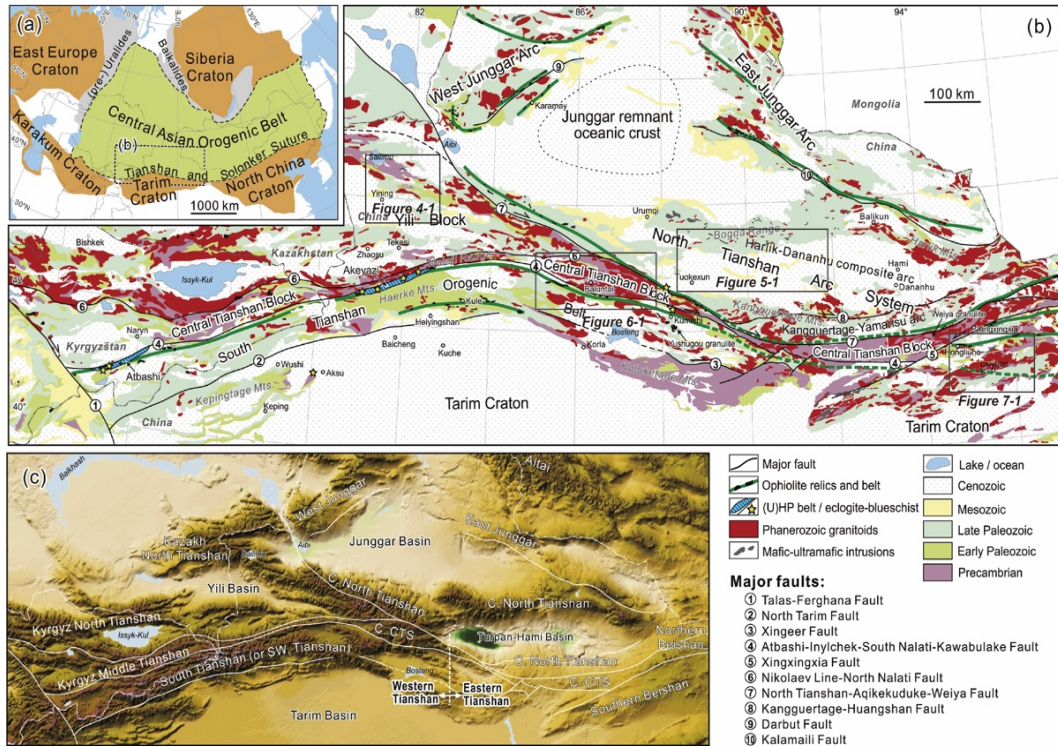


Figure 1. (a) Carte tectonique simplifiée de la CAO et des régions alentour. (b) Carte géologique du Sud de la CAO couvrant les ceintures de Tianshan et de Beishan. (c) Carte topographique des ceintures de Tianshan et Beishan (de Han et Zhao (2018)). Les localités de nos échantillonnages paléomagnétiques, couvrant les blocs de Yili, Baluntai et Turpan-Hami et la zone de Shuangyingshan, sont marquées sur Figure 1b.

Dans le bloc de Yili, des dykes basiques au flanc oriental du Lac de Sailimu et les roches volcaniques et sédimentaires d'âge permien du Goupe de Wulang dans les zones de Jinghe et Nileke étaient prélevés pour l'étude paléomagnétique. Les investigations de magnétisme des roches, y compris l'acquisition de l'aimantation rémanente isotherme (IRM), courbes d'expérimentation thermique, cycle de l'hystérésis, indiquent que la magnétite est le porteur principal de l'aimantation rémanente des dykes basiques du Néoprotérozoïque et la magnétite avec un peu d'hématite pour les roches volcaniques et sédimentaires du Permien. Le tuf andésitique échantillonné dans la zone de Jinghe présente un âge U-Pb sur zircon de 269.4 ± 2.4 Ma, ce qui confère un âge permien inférieur aux roches volcaniques et sédimentaires associées. Les aimantations

rémanentes caractéristiques (ChRMs) ont été isolées avec succès sur la majorité des échantillons en appliquant une désaimantation progressive par les techniques thermique et de champ alternatif (AF). Par des analyses raffinées sur les courbes de désaimantation, les roches volcaniques et sédimentaires d'âge permien inférieur portent probablement la rémanence magnétique primaire sur l'argument de leur test du pli positif. Par conséquent, un pôle paléomagnétique permien inférieur a été calculé pour le bloc de Yili à 81.5°N , 256.5°E , $A_{95} = 10.9^{\circ}$ avec $N = 11$ sites. Par contre, même si les dykes basiques néoproterozoïques ont donné un paléopôle magnétique très bien défini à 78.3°N , 211.8°E , $A_{95} = 7.2^{\circ}$ avec $N = 10$ sites, il est probable qu'ils ont été remagnétisés au Permien inférieur comme le suggèrent l'analogie de leurs directions de rémanence magnétique et les observations de terrain, en raison de l'intense flux thermique régional de cette époque. Toutefois, le caractère primaire de leur rémanence magnétique ne peut pas être totalement exclu sans de nouvelles données.

Des roches du Permien inférieur, principalement composées de grès fin/grossier et basalte dans les zones de Qijiaojing et Daheyan (marge nord du bassin de Turpan-Hami), ont été prélevées. Des analyses de magnétisme de roche, y compris l'expérimentation thermique, le cycle hystérésis et l'IRM, révèlent que ces échantillons peuvent être divisés en trois groupes d'après leurs minéraux magnétiques : hématite avec peu de magnétite ; magnétite avec peu d'hématite ; et magnétite. Après les corrections de pendage, les échantillons étudiés présentent une fabrique magnétique typiquement sédimentaire sans avoir subi de déformation ultérieure. La datation U-Pb sur zircon d'un échantillon de rhyolite et celle des zircons détritiques d'un grès de la partie inférieure de la séquence sédimentaire de Qijiaojing, ont fourni respectivement un âge à 310.5 ± 1.2 Ma pour la rhyolite et un âge maximum de dépôt du grès de ~ 294 Ma. En combinant avec les données géochronologiques disponibles, on peut estimer que l'âge de la séquence sédimentaire étudiée est permien inférieur. Sur la base du modèle TK03.GAD de la variation séculaire du champ géomagnétique, un affaiblissement significatif de l'inclinaison magnétique par la compaction a été identifié avec un facteur

d'aplatissement de 0.75 pour les sédiments de la zone de Qijiaojing ; par contre, ce phénomène n'a pas été observé dans la zone de Daheyan. Plusieurs arguments, dont le faible degré d'anisotropie, la représentation totale de la variation séculaire du champ géomagnétique par ChRMs, l'affaiblissement significatif de l'inclinaison de la séquence dans la zone de Qijiaojing, et l'existence d'une seule polarité inverse, suggèrent une rémanence primaire. Utilisant la correction élongation/inclinaison (E/I), le premier pôle paléomagnétique du Permien inférieur pour le bloc de Turpan-Hami a été obtenu : 75.7°N , 276.3°E , $A95 = 5.7^{\circ}$ avec $N = 10$ sites.

Une étude pluridisciplinaire a été effectuée sur la migmatite du bloc de Baluntai (Tianshan central chinois). Les échantillons sont séparés en trois groupes : leucosome, mésosome, et mélanosome. Sous le microscope, des structures typiquement magmatiques ont été observées dans les leucosome et mésosome, et un degré modéré de déformation dans le mélanosome. La datation U-Pb d'un granite migmatitique révèle un âge de 377.3 ± 5.1 Ma, et un échantillon de méta-greywacke présente un pic d'âge minimum de zircons détritiques à ~ 312 Ma. Les monazites du mélanosome révèlent des âges U-Th-Pb à 309-312 Ma. Par ailleurs, la datation $^{40}\text{Ar}/^{39}\text{Ar}$ sur des muscovites et biotites sélectionnées de granites migmatitiques et mélanosome présentent un plateau d'âge à 310.5 ± 6.5 Ma and 310.8 ± 7.6 Ma, respectivement. D'après les données d'ASM, les migmatites ont apparemment conservé une fabrique planaire pendant la migmatisation. Cependant, les ChRMs isolées des migmatites sont plutôt secondaires, l'âge de cette rémanence est probablement de ~ 310 Ma. Un pôle paléomagnétique fiable n'a pas pu être calculé en raison de l'incertitude sur l'attitude de la foliation migmatitique lors de l'acquisition de cette rémanence magnétique secondaire.

La première étude paléomagnétique a été menée sur des dykes doléritiques d'âge Paléozoïque supérieur- Mésozoïque inférieur dans la zone de Gupaoquan, région de Shuangyingshan au Sud de Beishan. L'expérimentation thermique et le cycle d'hystérésis montrent que la magnétite semble être le porteur de la rémanence magnétique de ces dykes avec un peu de pyrite, de pyrrhotite et d'hématite. L'étude de

l'AMS n'a pas révélé de déformation intense postérieure à l'intrusion, elle montre aussi des fabriques magnétiques distinctes entre les dykes et leurs roches encaissantes. Les observations par MEB montrent que les magnétites de forme irrégulière sont en poly-domaine et les titano-magnétites sont le porteur principal de la rémanence magnétique, aucun mineral authigène n'a été identifié. Le plagioclase pris sur la bordure d'un large dike a fourni des âges plateau $^{40}\text{Ar}/^{39}\text{Ar}$ à 260-284 Ma. Sur ces bases, y compris la morphologie de minéraux magnétiques, le comportement magnétique des échantillons, le faible degré de l'ASM, la solo polarité magnétique inverse, et le test du contact positif, la rémanence magnétique de ces dykes doléritiques pourrait être considérée comme primaire avec un âge d'aimantation à ~280 Ma. Deux scénarios avec et sans faible basculement des dikes ont été proposés sur la base des observations de terrain et de l'ASM. Par conséquent, le premier pôle paléomagnétique d'âge permien inférieur pour la ceinture orogénique de Beishan (BOB) a été calculé à 80.9°N, 299.2°E, $A_{95} = 5.4^\circ$, avec $N = 20$ dikes.

De cette étude, trois pôles paléomagnétiques de haute qualité ont été obtenus pour les blocs de Yili et Turpan, ainsi que pour le Sud de Beishan. Bien qu'un pôle paléomagnétique fiable n'ait pas pu être obtenu pour le bloc de Baluntai, notre étude pluridisciplinaire contribue à la compréhension préliminaire de l'évolution tectonique et la position paléogéographique de ce bloc au Paléozoïque.

Combinant ces données paléomagnétiques à celles déjà publiées sur la CAO, on peut tirer les conclusions suivantes :

(1) cinématique et signification géologique de zones de cisaillement à grande échelle

L'ensemble des pôles paléomagnétiques du Paléozoïque supérieur au Mésozoïque inférieur des blocs principaux de la chaîne du Tianshan (TOB) et des zones alentour, qu'ils soient publiés auparavant ou bien acquis pendant cette étude, se distribuent sur

un petit cercle centré sur le site étudié, ce qui suggère que les mouvements de ces blocs concernent essentiellement des rotations locales, tandis que les mouvements relatifs en latitude restent mineurs. Ces rotations relatives pourraient résulter de mouvements décrochants le long des failles séparant ces blocs.

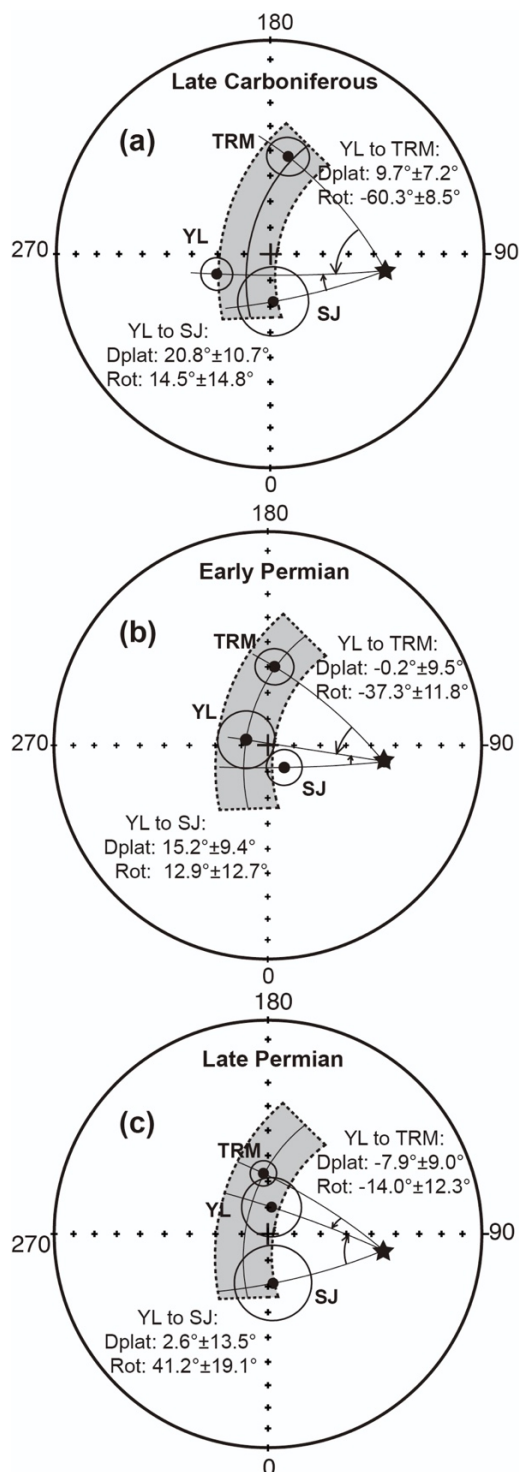


Figure 2. Projection équiaire (équivalente) des pôles paléomagnétiques des blocs de Yili (YL), Tarim (TRM), et le Sud de Junggar (SJ) au (a) Carbonifère supérieur, (b) Permien inférieur, et (c) Permien supérieur. Dplat et Rot représentent la convergence et rotation relative (avec les incertitudes), respectivement. Les valeurs positives (ou négatives) signifient un raccourcissement crustal (extension) ou rotation en sens horaire (antihoraire), respectivement.

Dans la chaîne du Tianshan (TOB), entre les blocs de Yili et Tarim, un mouvement dextre significatif a eu lieu le long d'une faille majeure (Nikolaiev Tectonic Line and Nalati Fault) du Carbonifère supérieur au Permien inférieur (580 ± 290 km) et du Permien inférieur au Permien supérieur (585 ± 340 km); La vitesse de déplacement augmente de $\sim 19.3 \pm 9.7$ mm/yr (Carbonifère - Permien

inférieur) à $\sim 29.3 \pm 17.0$ mm/yr (Permien inférieur – Permien supérieur) (Figure 2). Le mouvement relatif entre les blocs de Yili et Sud de Junggar est paléomagnétiquement négligeable pendant la période du Carbonifère au Permien inférieur ; cependant, une rotation relative de $28.3^\circ \pm 18.3^\circ$ au Permien supérieur et un déplacement latéral de 630 ± 295 km après le Permien supérieur ont eu lieu entre ces deux blocs le long de la faille dextre du Tianshan du Nord (Figure 2b-c).

Le bloc de Turpan-Hami a également subi des rotations relatives depuis le Permien inférieur de $-41.1^\circ \pm 7.1^\circ$ et $21.2^\circ \pm 9.4^\circ$ par rapport aux blocs du Tarim et Sud-Junggar, respectivement (Figure 3a), correspondant à des déplacements décrochants de grande ampleur le long de failles qui représentent le trait tectonique majeur de la partie SW de la CAOBS depuis le Carbonifère supérieur.

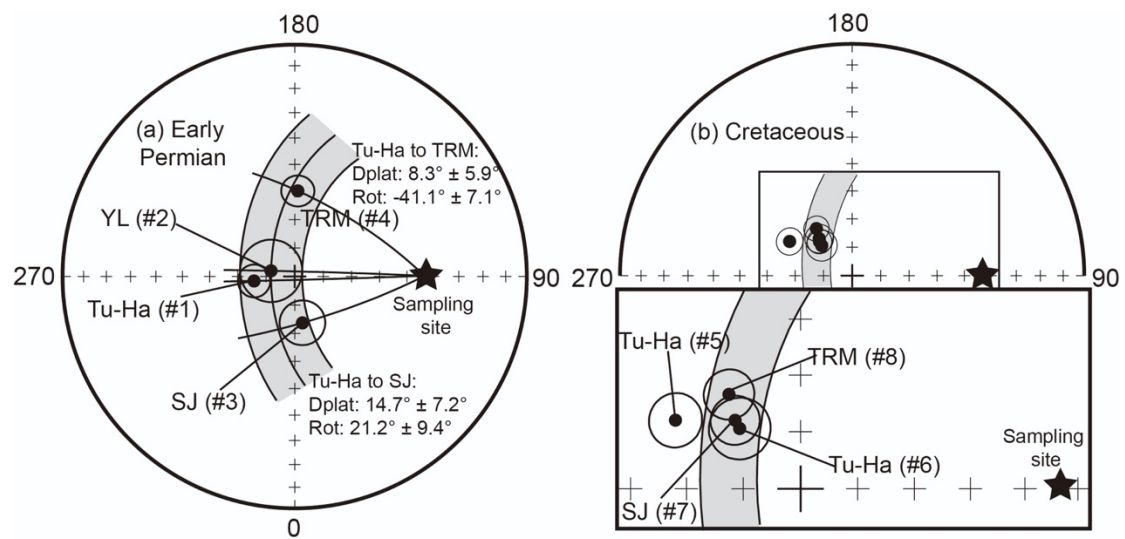


Figure 3. Projections équivalentes des pôles paléomagnétiques des blocs de Turfan-Hami (Tu-Ha), Yili (YL, Sud de Junggar (SJ) et Tarim (TRM) au (a) Permien inférieur et b) Crétacé.

Bien qu'un pôle paléomagnétique fiable ne puisse pas être établi pour le bloc de Baluntai en raison du manque de précision de la correction tectonique, les déclinaisons paléomagnétiques obtenues lors de notre étude sont encore significatives. Peu importe en coordonnées géologiques ou tectoniques (i.e., basculement de la foliation des

migmatites à l'horizontale), le bloc de Baluntai présente, en premier ordre, une déclinaison paléomagnétique similaire à celle du bloc du Tarim, mais il a subi des rotations importantes par rapport aux blocs de Yili, Turpan-Hami et, plus particulièrement, celui du Sud-Junggar depuis ~310 Ma.

Le pôle paléomagnétique d'âge permien inférieur du Sud de Beishan est proche de ceux de Yili, Turpan-Hami et Dunhuang, ce qui suggère que ces blocs se sont amalgamés depuis le Permien inférieur au moins et qu'il n'y a pas eu de mouvement relatif depuis lors. Par contre, les pôles paléomagnétiques du Permien inférieur des blocs de Tarim et du Sud-Junggar sont évidemment différents de celui de Beishan avec une rotation relative significative, qui révèle une implication similaire à ce qui a été mentionné auparavant. De telles rotations entre blocs ont été également observées dans la partie orientale du Sud de la CAO, cependant, leur importance doit être mieux contrainte.

(2) Transition de la collision/accrétion à la tectonique intracontinentale

Les études paléomagnétiques précédentes indiquent des différences négligeables en paléolatitudes entre les blocs majeurs dans la chaîne du Tianshan (TOB) depuis le Carbonifère supérieur. Notre étude confirme et raffine encore cette conclusion. Une différence angulaire de $6.6^\circ \pm 9.0^\circ$ a été calculée entre les pôles des blocs de Yili et Turpan-Hami pour le Permien inférieur (Figure 3a), suggérant qu'ils auraient atteint leur position actuelle depuis le Permien inférieur au moins. Une faible différence en paléolatitudes de $9.7^\circ \pm 7.2^\circ$ entre les blocs du Tarim et de Yili a été révélée au Carbonifère supérieur (Figure 2a) ; une faible différence en paléolatitudes de $8.3^\circ \pm 5.9^\circ$ a été aussi découverte entre les blocs de Turpan-Hami et Tarim au Permien inférieur (Figure 3a). Ces faibles différences en paléolatitudes soutiennent le modèle de l'extrusion vers l'est de certains blocs de la CAO. La ligne tectonique de Nikolaiev et la faille de Nalati ont une orientation NE-SW depuis le Paléozoïque supérieur, le faible mouvement vers le Nord des blocs de Yili et Turpan-Hami par rapport à celui de Tarim

pourrait être traduit par un simple mouvement décrochant le long de ces failles sans raccourcissement crustal. La fermeture du bassin extensif intra-arc d'une largeur de ~530 km entre les blocs de Turpan-Hami et Sud-Junggar pouvait être la cause du mouvement relatif significatif en latitude de $14.7^{\circ} \pm 7.2^{\circ}$ et de la surrection subséquente de Bogda depuis le Permien inférieur (Figure 3a). Cependant les données du Paléozoïque inférieur du Sud-Junggar proviennent de sites en faible nombre et ont donc besoin d'être réévaluées et testées avec de nouvelles données. Par ailleurs, les migmatites de Baluntai préservent une fabrique magnétique primaire acquise pendant la migmatisation, suggérant une période de relative stabilité tectonique liée probablement à la transition entre l'épaississement crustal et l'effondrement gravitaire post-orogénique au Carbonifère supérieur.

La simple comparaison de nos résultats avec les données publiées dans la région de Dunhuang, laisse apparaître une différence en paléolatitude de $7.0^{\circ} \pm 4.5^{\circ}$ d'âge permien inférieur entre Beishan et Dunhuang (Figure 4a). Cependant, il n'existe pas de preuve géologique de mouvement 'significatif' entre ces deux zones depuis le Permien inférieur, une analyse raffinée sur les données publiées de Dunhuang confirme ce doute.

Une différence en paléolatitude de $5.3^{\circ} \pm 7.2^{\circ}$ a été évoquée entre les blocs de Songliao-Xilinhote et de Chine du Nord à 265 Ma mais elle ne peut être considérée comme significative. En outre, des études paléomagnétiques récentes indiquent que le bloc de Mongolie est amalgamé avec le bloc de Chine du Nord depuis au moins le Permien inférieur. En conséquence, la transition de la collision/accrétion à la tectonique intracontinentale au Sud de la CAOBD date au moins du Permien inférieur.

(3) Reconstruction paléogéographique au Permien inférieur

D'après les données publiées et de celles de notre nouvelle étude, de négligeables différences en paléolatitude ont été détectées parmi les blocs majeurs soit dans la chaîne de Tianshan, Beishan ou bien la chaîne mongole depuis le Permien inférieur (Figure 4a), et ces blocs avaient leurs paléolatitudes décroissantes vers l'Est (au site de

référence : 41.0°N, 95.0°E) de 34°N à l'Ouest, 29°N au milieu, à 28° à l'Est. Gardant en tête la modification tardive de ce cadre géographique par des mouvements décrochants intracontinentaux, nous proposons une reconstruction paléogéographique plus plausible du Permien inférieur du Sud de la CAOBS avec le logiciel Gplates (Figure 4b).

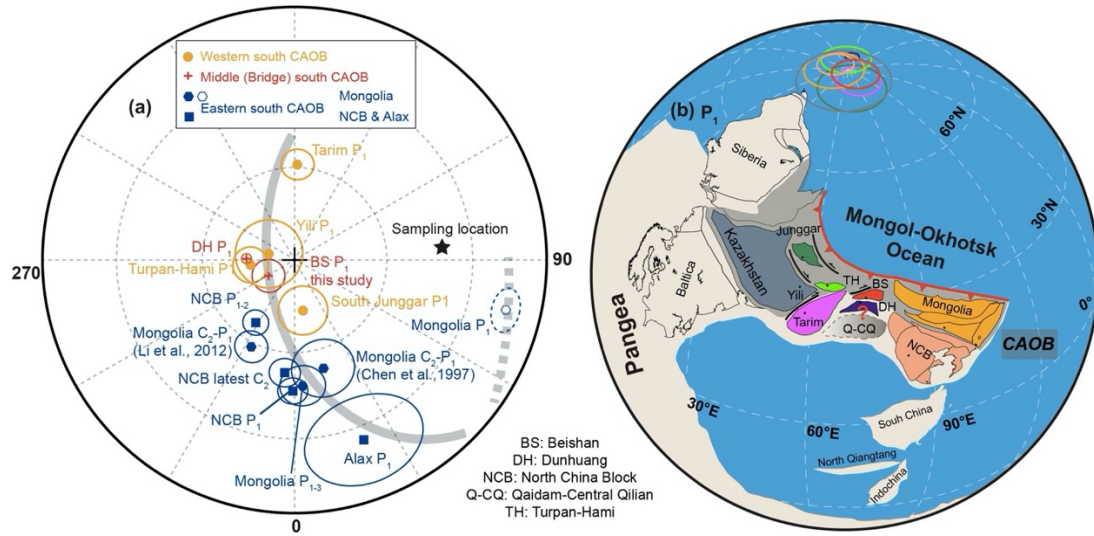


Figure 4. (a) Compilation des pôles paléomagnétiques d'âge permien inférieur des blocs majeurs au Sud de la CAOBS. (b) Reconstruction paléogéographique du Sud de la CAOBS au Permien inférieur.

En résumé, plusieurs pôles paléomagnétiques de haute qualité pour des microcontinents du Sud de la CAOBS ont été acquis pendant cette étude avec d'importantes implications tectoniques. De nouvelles techniques ont été mises en œuvre dans cette étude, telles que la correction E/I, le MEB couplé avec un spectroscope à dispersion d'énergie, la datation $^{40}\text{Ar}/^{39}\text{Ar}$ de divers minéraux, et l'application du logiciel Gplate pour la reconstruction paléogéographique, modernisant la procédure classique de l'étude paléomagnétique. Nos résultats paléomagnétiques fournissent une vision régionale et une démarche quantitative susceptible de résoudre les problématiques scientifiques mentionnées auparavant et approfondissent notre compréhension de l'évolution tectonique et paléogéographique du Sud de la CAOBS du

Paléozoïque au Mésozoïque inférieur.

Mots clés: Ceinture Orogénique d'Asie Centrale (CAOB), Paléozoïque Supérieur, Paléomagnétisme, Reconstruction Tectonique, Reconstruction Paléogéographique

Extended Abstract

The Central Asian Orogenic Belt (CAOB) or the Altaïds, located between the Siberia Craton to the north, East Europe (Baltica) Craton to the west, and the Tarim and North China cratons to the south, extending ~5500 km from the Urals in the west to the Okhotsk Sea in the east, is the largest Phanerozoic accretionary / collisional orogenic system on the Earth. After having attracted much attention from the geological community over several decades, it is well accepted now that the CAOB is a vast and composite orogenic system assembled by numerous microcontinents, seamounts, island arcs, and accretionary wedges through Neoproterozoic to Paleozoic successive multi-stage accretion and collision. Along with the final closure of the Paleo-Asian Ocean during late Paleozoic to early Mesozoic times, the CAOB was amalgamated via the docking of the North China and Tarim cratons along its southernmost part and experienced significant intracontinental reworking dominated by large-scale strike-slip faulting with horizontal displacements reaching a magnitude of several hundred kilometers. During the Mesozoic to Cenozoic, it was under the far-field effect of the collisions among several southern blocks, especially the India-Eurasia collision which resulted in extensive crustal shortening and uplift in the CAOB.

Numerous previous works have brought many important insights into the evolution of CAOB involving Neoproterozoic to Paleozoic multi-stage accretionary/collisional orogeny, late Paleozoic to early Mesozoic post-orogenic tectonic transition, and Mesozoic to Cenozoic intracontinental activated tectonics. However, due to its complexity and insufficient investigations, many controversies remain as to the late Paleozoic to early Mesozoic evolution. The southern part of the CAOB, covering the Tianshan Orogenic Belt, Beishan Orogenic Belt, and Mongolia Orogenic Belt, witnesses the subduction and closure of the Paleo-Asian Ocean and the final amalgamation with the Tarim and North China cratons. As such, it is of great significance in unveiling the tectonic evolution history and the orogen paleogeography

of the CAO B.

Major key issues that still need to be resolved in the southern CAO B for reaching a consensus can be identified as follows:

- Revealing the kinematics and geological significance of the large-scale strike-slip shear zones
- Understanding the tectonic transition from collision/accretion to intracontinental tectonics
- Achieving well-constrained late Paleozoic to early Mesozoic paleogeographic reconstructions

(1) Revealing the kinematics and geological significance of the large-scale strike-slip shear zones

According to paleomagnetic data, the magnitude of strike-slip displacements along the major dextral shear zones in the Tianshan Orogenic Belt reaches several hundred kilometers. These dextral shear zones coupled with the Erqis sinistral shear zone indicate an eastward extrusion of the Yili, Junggar, and North Tianshan units relative to the Siberia and Tarim cratons, which probably resulted from intraplate oblique convergence. Large-scale strike-slip shear zones also developed in the Beishan Orogenic Belt and Mongolia Orogenic Belt but are less investigated. Recent paleomagnetic studies in the Mongolia Orogenic Belt suggest that significant post-Early Permian vertical-axis strike-slip-related rotations occurred within and along the margins of the unified Mongolia Block and North China Craton, but their kinematic senses and magnitudes are not further constrained.

Many scholars regard these strike-slip displacements as the intracontinental adjustment after the closure of the Paleo-Asian Ocean. Indeed, strike-slip faults across the southern CAO B are well-developed and can form a west-east fault system with basically similar ages and kinematics. Such a fault system hints that the large-scale

strike-slip movement is the main intracontinental tectonic pattern after the amalgamation and must have played an important role in building the final configuration of the southern CAOB, forming accompanied late Paleozoic to early Mesozoic magmatism and mineral deposits, and deciphering paleo-geodynamics in this region. Moreover, these shear zones provide an excellent opportunity to study stress partitioning and strain localization in regional-scale transpressional and transtensional deformation zones. Therefore, more studies are requisite in revealing the kinematics and geological significance of the large-scale strike-slip shear zones in the southern CAOB.

(2) Understanding the tectonic transition from collision/accretion to intracontinental tectonics

How the tectonic transition, from collision/accretion to intracontinental tectonics, performed remains unclear. The timing, pattern, and geodynamic process of this transition have long been the focus issues of academic controversy. Different final closure time of the Paleo-Asian Ocean have been proposed in and between its different segments. Several hypotheses, including oblique convergence, slab break-off, and extrusion tectonics, have been proposed to interpret the end Paleozoic to early Mesozoic tectonic transition. Furthermore, different tectonic settings were also suggested for the late Paleozoic tectonic and magmatic environments, such as subduction or ridge subduction, post-orogenic extension/rifting, and super mantle plume-related large igneous province. The consensus is difficult to reach on the above issues, mainly because of limited previous research approaches (chiefly geochronology and geochemistry of igneous rocks) that sometimes lead to ambiguous tectonic interpretations. Therefore, more constraints from different perspectives are greatly needed.

Kinematics along the major boundaries between microcontinental blocks and terranes in the southern CAOB during late Paleozoic to early Mesozoic time is the key

to decipher the geodynamic evolution and understand mechanisms of transition from subduction/accretion to intracontinental tectonics. However, these boundaries experienced long and multi-stage deformation during orogenic processes; therefore, geodynamic reconstruction using geological records becomes complicated and controversial. Paleomagnetism is a workable method to tackle this issue, as it provides quantitative, large-scale constraints on the kinematics of blocks, including paleolatitudinal translation and vertical-axis rotation. Furthermore, through comparisons of paleomagnetic poles of different tectonic units in the southern CAOB, the kinematic evolution, in terms of velocity, orientation, and pattern, among these tectonic units during a specific geological period could be revealed, providing a crucial basis for understanding such tectonic transition.

(3) Achieving well-constrained late Paleozoic to early Mesozoic paleogeographic reconstructions

In the attempt to reconstruct the configuration of the supercontinent Pangea, scholars found that the main compositions (large cratons) of Pangea are well framed through voluminous evidence from robust paleomagnetic, geological, and paleontological data. However, the reconstruction of the East Asian blocks in Pangea remains controversial. The leading cause is that the geological structure in East Asian is complicated, especially in the CAOB, as this latter is composed of numerous and variously deformed microcontinents, island arcs, seamounts, and accretionary wedges. Due to this complexity, few reliable paleomagnetic data had been obtained from the microcontinents (except the Tarim Block) in the CAOB, which can crucially constrain the paleogeographic framework. As a result, no robust paleogeographic reconstruction for the CAOB had been achieved.

However, paleogeography is essential to comprehend the origin and diversity of life, ancient climate, and the distribution of tectonic belts and mineral resources. More specifically, the relative paleogeographic positions of different tectonic units in the

southern CAOBS are an essential foundation for the restoration and reconstruction of its paleo-ocean-continent framework, the evolution of the oceanic basin(s) and continental drift, and related orogenic processes. Recent research progress in the Himalayan Orogenic Belt is a good analogy. Thus, well-constrained late Paleozoic to early Mesozoic paleogeography of the southern CAOBS is significantly required, which in turn demands more high-quality paleomagnetic works on the microcontinents/terranes in the southern CAOBS.

Obviously, high-quality paleomagnetic data for the microcontinents/terranes in the southern CAOBS play a crucial role in working out the above scientific problems. Aiming to provide more constraints on the late Paleozoic to early Mesozoic evolution of the southern CAOBS, we have conducted paleomagnetic studies coupled with field observation, thin-section observation, rock magnetism, the Anisotropy of Magnetic Susceptibility (AMS) analysis, scanning electronic microscopic (SEM) imaging, geochronology, and thermochronology on the Neoproterozoic and Early Permian mafic dikes, late Carboniferous migmatites, and Early Permian volcanic and sedimentary rocks collected from the Yili, Turpan-Hami, and Baluntai blocks in the Tianshan Orogenic Belt, as well as the Shuangyingshan terrane in the Beishan Orogenic Belt (Figure 1).

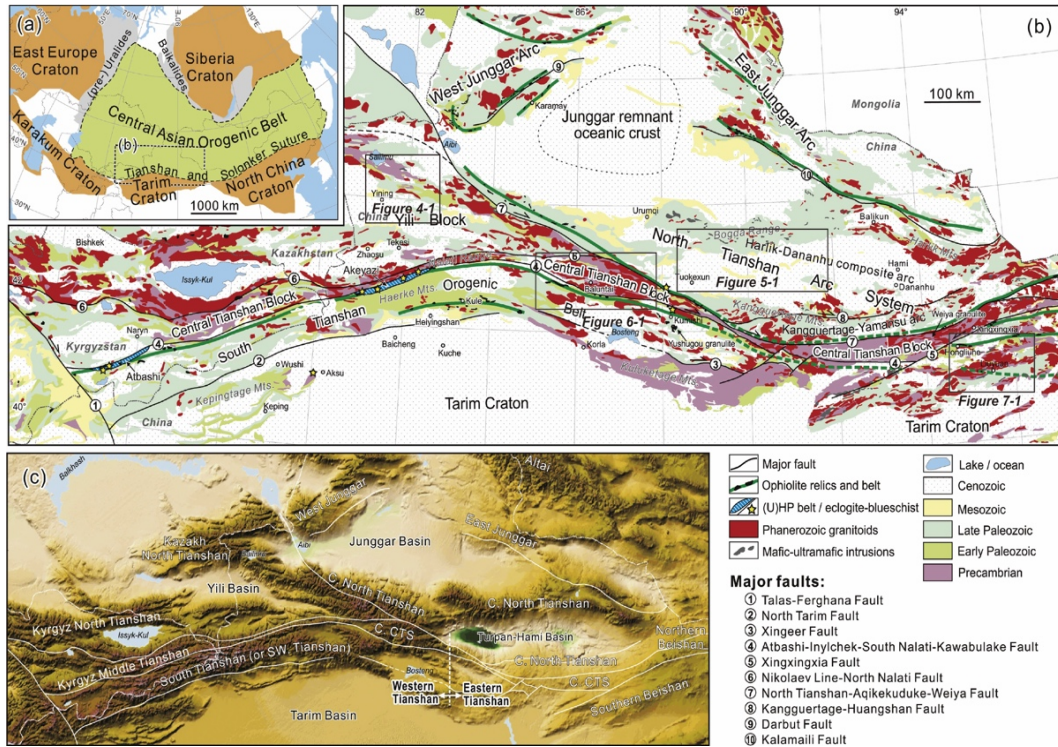


Figure 1. (a) Tectonic sketch of the CAOB and its surrounding regions. (b) Geological map of the southern CAOB covering the Tianshan and Beishan belts. (c) Topographic map of the Tianshan and Beishan belts (from Han and Zhao (2018)). Our paleomagnetic sampling locations, covering the Yili, Baluntai, and Turpan-Hami blocks and Shuangyingshan terrane, are marked on the Figure 1b.

In the Yili Block, Neoproterozoic mafic dikes in the eastern side of the Sailimu Lake and Early Permian volcanic and sedimentary rocks of the Wulang Group in the Jinghe and Nileke areas were sampled for the paleomagnetic study. Rock magnetic investigations, including Isothermal Remanent Magnetization (IRM) acquisition curves, thermomagnetic experiment, and hysteresis loops, indicated that magnetite is the main magnetic mineral of the Neoproterozoic mafic dikes and that magnetite with minor hematite is the magnetic minerals of the Early Permian volcanic and sedimentary rocks. The andesitic tuff sample collected from the Jinghe area shows a U-Pb zircon age of 269.4 ± 2.4 Ma, suggesting that the studied volcanic and sedimentary rocks were formed in the Early Permian. Characteristic remanent magnetizations (ChRMs) were

successfully isolated from most samples using stepwise thermal and alternating field demagnetizations. Through careful analyses on the demagnetization curves, the Early Permian volcanic and sedimentary rocks probably carry primary magnetic remanence argued by their positive fold tests. Therefore, an Early Permian paleomagnetic pole for the Yili Block is obtained at 81.5°N , 256.5°E , $A_{95} = 10.9^{\circ}$ with $N = 11$ sites. In contrast, although an Early Permian remagnetization is preferred for the Neoproterozoic mafic dikes based on their magnetic remanence directions, field observation, and regional magmatic activities, a primary origin cannot be ruled out until more data are obtained. Therefore, further work is needed to interpret the paleomagnetic pole acquired from the mafic dikes: 78.3°N , 211.8°E , $A_{95} = 7.2^{\circ}$ with $N = 10$ sites.

Lower Permian strata, mainly consisting of fine/coarse-grained sandstones and basalts, from both Qijiaojing and Daheyan areas (northern margin of Turpan-Hami basin) were collected. Rock magnetic mineralogic analyses, including thermomagnetic experiments, hysteresis measurements, and IRM acquisition, revealed that the studied samples could be divided into three groups according to their major magnetic minerals: hematite with minor magnetite; magnetite with minor hematite; and magnetite. After tilt corrections, the studied samples show a typical magnetic fabric related to the depositional process without intensive posterior deformation. Zircon U-Pb dating of two geochronological samples, one rhyolitic cobble and one sandstone from the lower part of the sampled Qijiaojing sequence, provides a weighted mean age at 310.5 ± 1.2 Ma and a maximum depositional age at ~ 294 Ma, respectively. Combining with published geochronological data, the formation age of the studied sequence is verified as Early Permian. Based on the paleosecular variation model TK03.GAD of the geomagnetic field, significant magnetic inclination shallowing with a flattening factor $f = 0.75$ was found for the strata in the Qijiaojing area; in contrast, no apparent magnetic inclination shallowing occurs in the Daheyan strata. Several lines of evidence, including low anisotropy degree, a full representation of the paleosecular variation of the geomagnetic field by ChRMs, significant inclination shallowing of the strata in the Qijiaojing area,

and sole reverse polarity of ChRMs, all support a primary remanence. Applying elongation/inclination (E/I) correction, the first Early Permian paleomagnetic pole for the Turpan-Hami Block was estimated at 75.7°N , 276.3°E , $A95 = 5.7^{\circ}$, and $N = 10$ sites.

A multidisciplinary study has been performed on the migmatite in the Bailuntai Block (Chinese Central Tianshan). Samples were separated into three groups: leucosome, mesosome, and melanosome. Under a binocular microscope, typical magmatic structures were found in the leucosome and mesosome, while a moderate degree of deformation was observed in the melanosome. Zircon U-Pb dating of a migmatitic granite provides a crystallization age at 377.3 ± 5.1 Ma, and a meta-greywacke sample shows the youngest U-Pb detrital zircon age peak at ~ 312 Ma. Monazite selected from melanosome presents U-Th-Pb ages at 309-312 Ma. Moreover, $^{40}\text{Ar}/^{39}\text{Ar}$ dating performed on muscovite and biotite selected from migmatitic granite and melanosome show plateau ages over all the steps at 310.5 ± 6.5 Ma and 310.8 ± 7.6 Ma, respectively. Based on the AMS data, the migmatite likely preserves a “primary” horizontal magnetic fabric acquired during the migmatization. However, the ChRM isolated from the migmatite is a secondary one, whose remanence age is probably ~ 310 Ma. No reliable paleomagnetic pole can be obtained so far because the attitude of migmatitic foliation during the magnetic remanence acquisition is unknown.

The first late Paleozoic to early Mesozoic paleomagnetic study for the Beishan Orogenic Belt was carried out on the Early Permian doleritic dikes in the Gubaoquan area, Shuangyingshan terrane in south Beishan. Thermomagnetic experiment and hysteresis loops revealed that magnetite seems the principal magnetic mineral of the doleritic dikes with some pyrite, pyrrhotite, and hematite. AMS study showed no intensive deformation of the mafic dikes and different magnetic fabrics between the dikes and their country rocks. SEM observation identified that irregular multi-domain magnetite and titanomagnetite are the main magnetic remanence carriers, and no authigenic mineral is found. The plagioclase collected from the margin of a wide doleritic dike gave $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages at 260-284 Ma. As supported by a series of

evidence, including magnetic minerals' morphologies, dike specimens' magnetic behaviors, low anisotropy degree, single reversed polarity, and a baked contact test, and so on, the magnetic remanence carried by the mafic dikes is probably primary with a remanence age at ~280 Ma. Moreover, a non/weak-tilting scenario is proposed based on field observation and AMS data. Accordingly, the first Early Permian paleomagnetic pole for the Beishan Orogenic Belt is acquired at 80.9°N, 299.2°E, $A95 = 5.4^\circ$, and $N = 20$ dikes.

From this Ph.D. study, three new high-quality Early Permian paleomagnetic poles for the Yili and Turpan-Hami blocks and south Beishan have been obtained. Although no reliable paleomagnetic pole for the Baluntai Block can be provided, our multidisciplinary study enables us to reach a preliminary understanding of the tectonic evolution and paleogeographic position of the Baluntai Block in the late Paleozoic.

Combining with published paleomagnetic data from the southern CAOB, the following conclusions can be drawn:

(1) Kinematics and geological significance of the large-scale strike-slip shear zones

Previous and our newly acquired late Paleozoic to early Mesozoic paleomagnetic poles for the major blocks in the Tianshan Orogenic Belt and surrounding areas distribute along a small circle, which suggests that the relative movements among these blocks are dominated by relative rotations while relative latitudinal movements are subordinate. These relative rotations could be interpreted by the significant strike-slip movements along the major faults among these blocks.

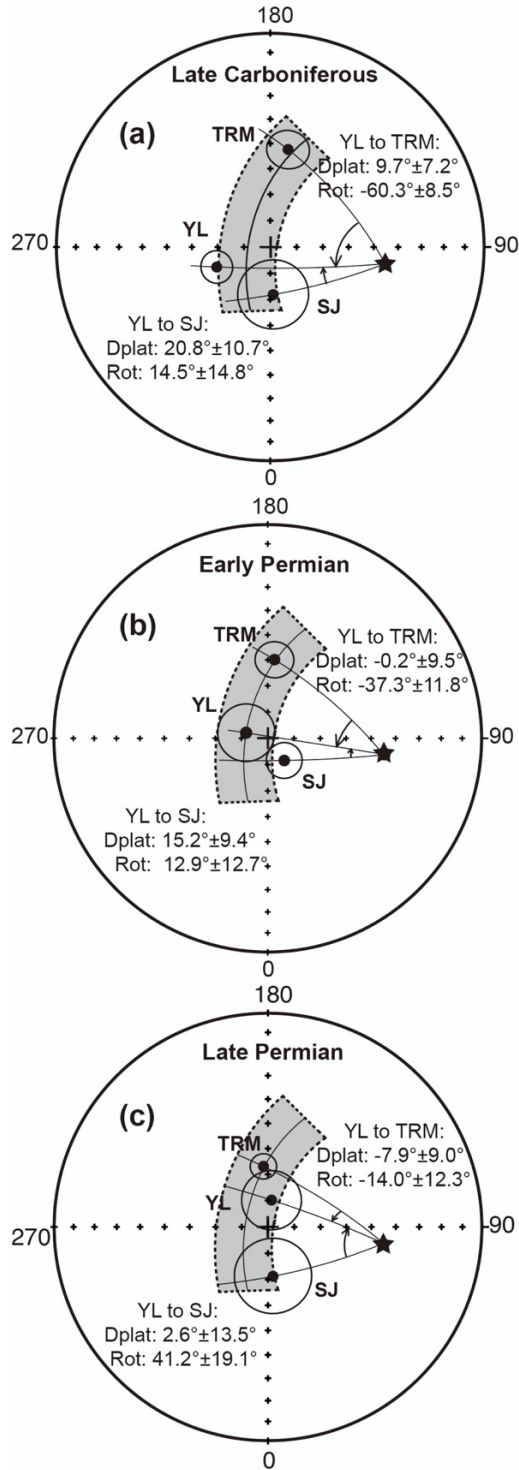


Figure 2. Equal-area projections of paleomagnetic poles of the Yili (YL), Tarim (TRM), and South Junggar (SJ) blocks in (a) Late Carboniferous, (b) Early Permian, and (c) Late Permian. Dplat and Rot stand for convergence and relative rotation (with corresponding uncertainties), respectively. The positive (or negative) mean values stand for latitudinal shortening (or extension) or clockwise (counterclockwise) rotation, respectively.

In the Tianshan Orogenic Belt, between the Yili and Tarim blocks, a significant relative movement took place along major dextral strike-slip faults (Nikolaiev Tectonic Line and Nalati Fault) during the Late Carboniferous to Early Permian (580 ± 290 km) and the Early to Late Permian (585 ± 340 km); the displacement rate increased from the Late Carboniferous to Early Permian ($\sim 19.3 \pm 9.7$ mm/yr) to the Early to Late Permian ($\sim 29.3 \pm 17.0$ mm/yr) (Figure 2). Relative movement is paleomagnetically negligible between the Yili and South Junggar blocks during the Late Carboniferous to Early Permian, but a significant relative rotation of $28.3^{\circ} \pm 18.3^{\circ}$ in the Late Permian and lateral displacement of 630 ± 295 km after the Late Permian occurred between these two blocks along the dextral North Tianshan Fault

(Figure 2b-c).

The Turpan-Hami Block also experienced relative rotations of $-41.1^\circ \pm 7.1^\circ$ and $21.2^\circ \pm 9.4^\circ$ with respect to the Tarim and South Junggar blocks since the Early Permian, respectively (Figure 3a), corresponding to the large-magnitude strike-slip displacements along major strike-slip faults characterizing the main intracontinental tectonic features of the SW CAOB since the late Carboniferous.

Although no reliable paleomagnetic pole could be given for the Baluntai Block due to a lack of tectonic correction, its magnetic declination obtained from our study is still meaningful. No matter in the geological coordinates or the tectonic coordinates (i.e., tilting the migmatitic foliation into horizontal), at the first-order, the Baluntai Block show similar magnetic declination with respect to the Tarim Block, while it experienced large-scale clockwise rotations with respect to the Yili, Turpan-Hami blocks and especially the South Junggar Block since ~ 310 Ma.

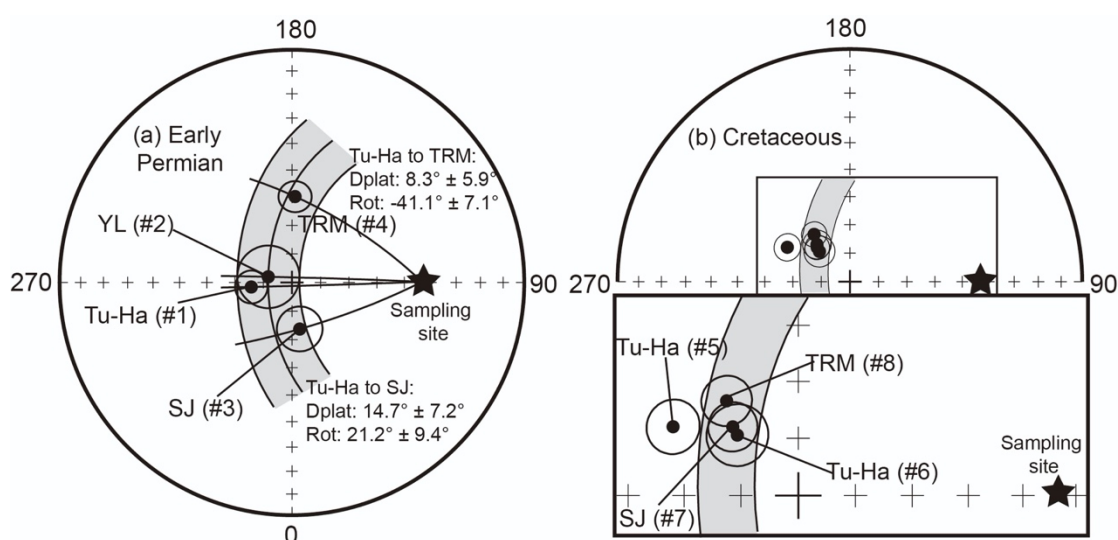


Figure 3. Equal-area projections of paleomagnetic poles of the Turpan-Hami (Tu-Ha), Yili (YL), South Junggar (SJ) and Tarim (TRM) blocks in (a) the Early Permian and (b) Cretaceous.

The Early Permian paleomagnetic pole of south Beishan is close to those of Yili, Turpan-Hami, and Dunhuang blocks, suggesting that these blocks were amalgamated

together at least since the Early Permian and no paleomagnetically significant relative movement occurred among them thereafter. In contrast, the Early Permian paleomagnetic poles of the Tarim and South Junggar blocks are obviously different from that of south Beishan with significant relative rotation, which has a similar geological implication as aforementioned. Actually, significant relative rotations among major blocks are also presented in the eastern part of the southern CAOB, but their significance needs more investigations.

(2) Tectonic transition from collision/accretion to intracontinental tectonics

Previous paleomagnetic studies indicated no significant latitudinal difference between major blocks in the Tianshan Orogenic Belt since the late Carboniferous. Our study confirmed and further refined this conclusion. An indistinguishable angular difference of $6.6^{\circ} \pm 9.0^{\circ}$ is found between the Early Permian paleomagnetic poles of the Yili and Turpan-Hami blocks (Figure 3a), suggesting that they reached similar relative positions as today at least in the Early Permian. A weak latitudinal difference of $9.7^{\circ} \pm 7.2^{\circ}$ is revealed between the Yili and Tarim blocks in the Late Carboniferous (Figure 2a); a weak latitudinal difference of $8.3^{\circ} \pm 5.9^{\circ}$ is also discovered between the Turpan-Hami and Tarim blocks in the Early Permian (Figure 3a). These weak latitudinal differences further support the eastward extrusion model. Since the Nikolaiev Tectonic Line and Nalati Fault strike NE-SW in the late Paleozoic, the weak northward movement of the Yili and Turpan-Hami block with respect to the Tarim Block might occur as a simple “sliding” along the fault instead of crustal shortening. The closure of a ~530 km-width intra-arc rift basin existed between the South Junggar and Turpan-Hami blocks in the Early Permian and subsequent uplift of the Bogda Range can be used to interpret the significant latitudinal difference of $14.7^{\circ} \pm 7.2^{\circ}$ between the Turpan-Hami and South Junggar blocks (Figure 3a). However, the late Paleozoic paleomagnetic data for the South Junggar Block are from few sites and need to be evaluated and tested. In addition, the Baluntai migmatite preserves a “primary”

horizontal magnetic fabric acquired during the migmatization, indicating a relatively stable tectonic stage probably related to the transition from orogenic thickening of crust to the post-orogenic gravitational collapse in the late Carboniferous.

The simple comparison of our results with the data published in the Dunhuang region, reveals a difference in paleolatitude of $7.0^{\circ} \pm 4.5^{\circ}$ of Early Permian age between Beishan and Dunhuang (Figure 4a). However, there is no geological evidence of "significant" movement between these two areas since the Early Permian, and a refined analysis of published data from Dunhuang confirms this doubt.

An insignificant latitudinal difference of $5.3^{\circ} \pm 7.2^{\circ}$ was also reported between the Songliao-Xilinhote Block and North China Craton at 265 Ma. Furthermore, recent paleomagnetic work indicates the Mongolia Block should have been welded with the North China Craton at least in the Early Permian. Consequently, the tectonic transition from collision/accretion to intracontinental tectonics in the southern CAOB was probably completed at least in the Early Permian.

(3) Early Permian paleogeographic reconstruction

According to previous and our own paleomagnetic data, no noticeable latitudinal difference is detected among major blocks either in the Tianshan, Beishan, or Mongolia orogenic belts since the Early Permian and the major blocks have Early Permian paleolatitudes decreasing eastwards (reference site: 41.0°N , 95.0°E) from $\sim 34^{\circ}\text{N}$ (west), $\sim 29^{\circ}\text{N}$ (middle) to $\sim 28^{\circ}\text{N}$ (east). Keeping in mind the significant later modification by large-scale intracontinental strike-slip movements, we propose a plausible Early Permian paleogeographic reconstruction of the southern CAOB using software Gplates (Figure 4b).

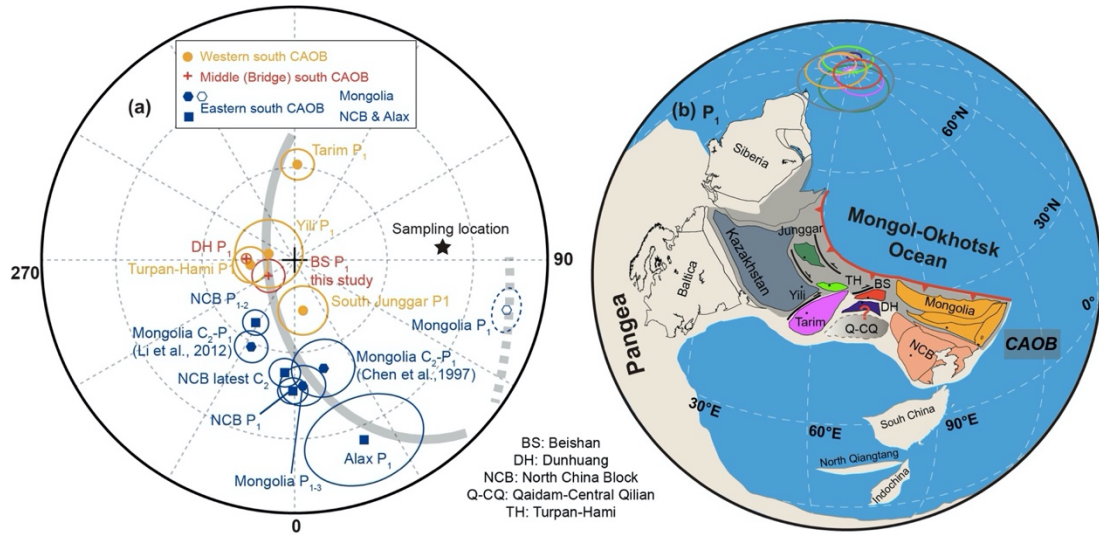


Figure 4. (a) Compilation of Early Permian paleomagnetic poles for major blocks in the southern CAOB. (b) Early Permian paleogeographic reconstruction of the southern CAOB.

In summary, several high-quality Early Permian paleomagnetic poles for the microcontinents in the southern CAOB have been acquired during our study, which are of great geological implication but have been lacking for a long time. New techniques, such as E/I correction, SEM observation coupled with the energy-disperse spectroscope, multi-minerals $^{40}\text{Ar}/^{39}\text{Ar}$ dating, and paleogeographic reconstruction software Gplates, were used in this study, further developing the traditional paleomagnetic study process. Our paleomagnetic results provided a regional view and a quantitative way to resolve the abovementioned remaining scientific problems in the southern CAOB and reach a deeper understanding of the late Paleozoic to early Mesozoic evolution and paleogeography of the CAOB.

Key words: Central Asian Orogenic Belt (CAOB), Late Paleozoic, Paleomagnetism, Tectonic Reconstruction, Paleogeographic Reconstruction

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Chapter 1. General Introduction

1.1 Research status in the southern Central Asian Orogenic Belt

Located between the Siberia Craton to the north, East Europe (or Baltica) Craton to the west, and the North China, Tarim, and Karakum cratons to the south, the Central Asian Orogenic Belt (CAOB; Figure 1-1; Yanshin et al., 1965; Jahn, 2004) or the Altaïds (Sengör et al., 1993; Xiao et al., 2010), extending ~5500 km from the Okhotsk Sea in the east to the Urals in the west, is the largest Phanerozoic accretionary/collisional orogenic system on the Earth. It is widely accepted that the CAO B is a composite and vast orogenic system assembled by numerous microcontinents, seamounts, island arcs and accretionary wedges through Neoproterozoic to Paleozoic successive accretion and collision (e.g., Windley et al., 2007; Levashova et al., 2009, 2011; Xiao et al., 2010, 2013; Geng et al., 2011; Wilhem et al., 2012; Eizenhöfer et al., 2014; Han and Zhao, 2018). As a result of the closure of the Paleo-Asian Ocean during late Paleozoic to early Mesozoic time, the CAO B was amalgamated via the collision of the North China and Tarim cratons along its southernmost part (Han and Zhao, 2018). Since late Paleozoic, the CAO B experienced significant intracontinental reworking dominated by large-scale strike-slip faulting (e.g., Allen et al., 1993; Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2009; Zhao et al., 2015, 2020). Afterwards, multiple uplifting events occurred in the CAO B. During the Cenozoic, it underwent extensive crustal shortening and uplift due to the far-field effect of the India-Eurasia collision (e.g., Molnar and Tapponnier, 1975; Chen et al., 1992; Avouac et al., 1993a, b).

The late Paleozoic to early Mesozoic evolution, mainly in terms of paleogeography and tectonics, of the CAO B is of great geological importance, as it witnesses the ultimate amalgamation of the CAO B along with the final closure of the Paleo-Asian Ocean (Eizenhöfer and Zhao, 2018; Han and Zhao, 2018; Zhou et al., 2018), and the

tectonic transition from multi-stage collision/accretion to large-scale strike-slip faulting-dominated intracontinental tectonics as well (Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2009). Moreover, during late Paleozoic to early Mesozoic, although the configuration of the main body of supercontinent Pangea is widely accepted, the reconstructions of East Asian blocks are still controversial (Zhao et al., 2018). The paleogeographic and tectonic reconstruction of the CAOAB during this geological period will shed light on settling this controversy.

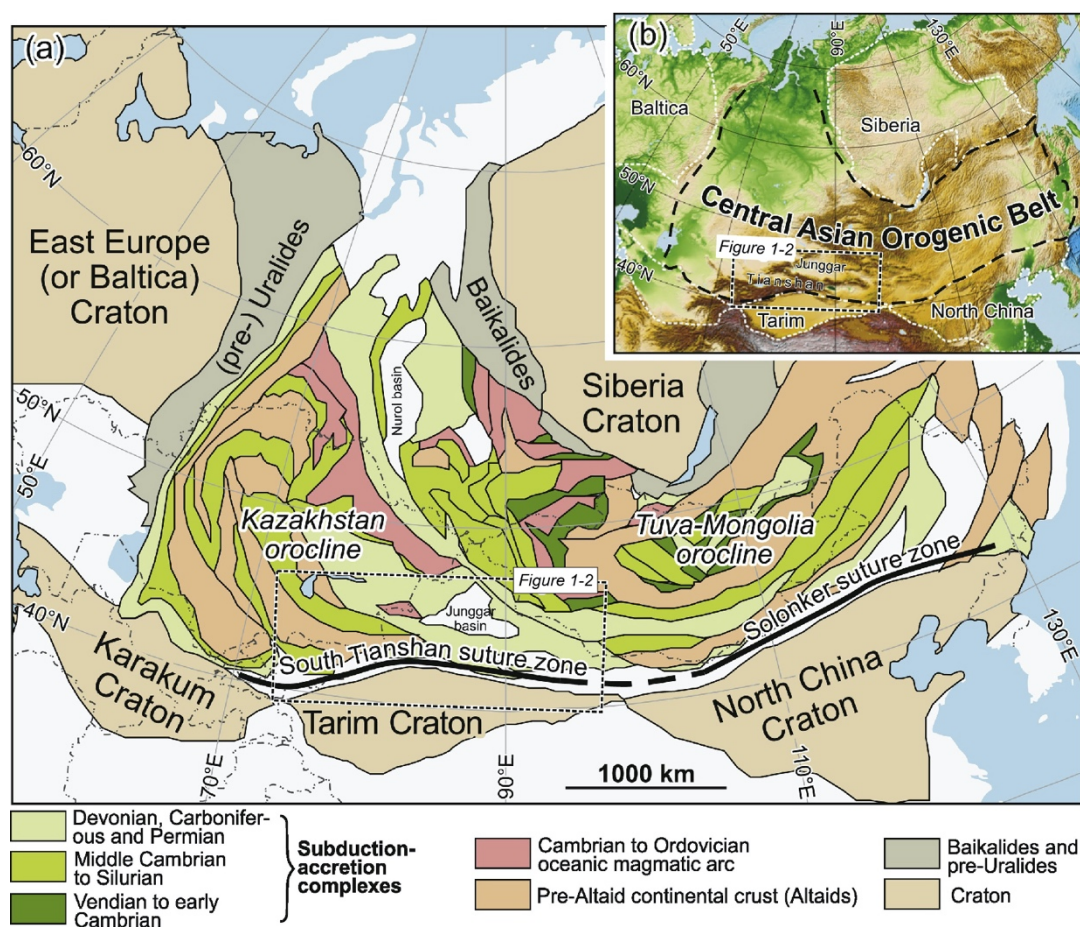


Figure 1-1. (a) Tectonic overview and division of the Central Asian Orogenic Belt and adjacent cratons. (b) Topographic features of Central Asia (from Han and Zhao (2018)). The southernmost part of the CAOAB can be divided into the Tianshan Belt, Beishan Belt, and Mongolia Belt. Figure 1-2 shows sampling area in this study.

Over the past decades, numerous studies have been carried out on the late Paleozoic to early Mesozoic evolution of the CAOAB. Especially, the southern part of the CAOAB

received intensive attention since it is the key region for deciphering aforementioned evolution process. In the following, current research status is summarized on the southern part of the CAO, which can be subdivided into the Tianshan Belt, Beishan Belt, and Mongolia Belt (Xiao et al., 2010).

1.1.1 The Tianshan Belt

The Tianshan Belt represents an orogenic belt formed by the collision/accretion of several microcontinental blocks, island arcs, and accretionary complexes, through the closure of the west segment of the Paleo-Asian Ocean, i.e. the South Tianshan Ocean and the North Tianshan Ocean (or Junggar Ocean).

Some scholars suggest that the South Tianshan Ocean closed earlier in the Late Devonian to earliest Carboniferous mainly based on the regional unconformities in the South Tianshan region, and the final formation of the Tianshan Belt resulted from the closure of the North Tianshan Ocean (Allen et al., 1993; Shu et al., 2002; Wilhem et al., 2012). According to Middle Devonian to early Carboniferous fossil, including radiolarians and conodonts, and radiometric ages from the ophiolite relics (Liu, 2001; Shu et al., 2007; Wu and Li, 2013), (ultra)high-pressure metamorphic records in the Akeyazi belt and Atabashi belt, (Gao and Klemd, 2003; Hegner et al., 2010; Klemd et al., 2011), magmatic records (Yuan et al., 2010; Qin et al., 2011), and large-scale strike-slip displacements (Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2009), other authors propose a roughly contemporaneous closure of both South Tianshan and North Tianshan Oceans in the late Carboniferous (Gao et al., 2011; Klemd et al., 2015; Han and Zhao, 2018). However, some authors raise a late Permian to mid-Triassic final closure of the western Paleo-Asian Ocean (Windley et al., 2007; Xiao et al., 2010, 2013), on account of the late Permian radiolarians from the Qiqijianake ophiolite (Li et al., 2002, 2005), and oceanic subduction-related magmatism (Mao et al., 2014).

Several roughly E-W extending large-scale shear zones developed in the Tianshan area, including Nikolaev line (Burtman et al., 1975), Nalati Fault (Gao et al., 2009),

North Tianshan Fault (Zhou et al., 2001; Wang et al., 2006, 2009) and Main Tianshan Shear Zone (Laurent-Charvet et al., 2002, 2003). Many structural and geochronological studies have been conducted on these shear zones, finding that they generally have dextral shear sense and late Carboniferous to latest Permian $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages (Laurent-Charvet et al., 2002, 2003; Wang et al., 2008b; de Jong et al., 2009; details in the section 2.1.2). Such regional strike-slip shear zones likely reflect an intracontinental adjustment after the final amalgamation of the Tianshan Belt (Wang et al., 2007; Choulet et al., 2011; Yi et al., 2015).

Paleomagnetism, as the universal quantitative method for conducting paleogeographic and kinematic reconstructions of the continental plates, is a powerful technique to understand paleogeographic configuration, tectonic transition, and the amounts of the large-scale strike-slip displacements of the Tianshan Belt. Available paleomagnetic data in the Tianshan area focus on the Tarim, Yili, and Junggar blocks. Abundant paleomagnetic works have been carried out on the Tarim Block from the Neoproterozoic to the Cenozoic (e.g. Zhao et al., 2014; references therein and more information in Chapter 8). Conducting the first paleomagnetic study on the volcanic and sedimentary rocks from the Yili Block, Wang et al. (2007) provided the first Late Carboniferous and Late Permian paleomagnetic poles for the Yili Block, showing insignificant latitudinal difference between the Tarim, Yili, Junggar, and Siberia blocks since the late Carboniferous; however, large scale strike-slip displacements with several hundred kilometers magnitude along major faults occurred since then. Choulet et al. (2011, 2013) performed paleomagnetic works on the Junggar block and reached basically the same conclusions; moreover, they suggested that the Junggar Block was not a rigid block until the end of the Paleozoic.

1.1.2 The Beishan Belt

As the middle segment of the southern CAOB, the Beishan Belt links the Tianshan Belt to the west and the Mongolia Belt to the east. Traditionally, ophiolites and

ophiolitic mélanges are divided into four belts, namely Liuyuan Unit, Hongliuhe-Xichangjing, Jijitaizi-Xiaohuangshan, and Hongshishan ophiolitic mélanges from south to north (Xiao et al., 2010; section 2.2). The Hongliuhe-Xichangjing belt represents a Paleozoic ocean lasted at least from ~540 Ma to ~414 Ma (Cleven et al., 2015; section 2.2.4). Zircon U-Pb ages of the mafic rocks from the Jijitaizi-Xiaohuangshan belt show Carboniferous formation ages from ~345 Ma to ~321 Ma (section 2.2.6). Wang et al. (2014c) reported the first high-quality U-Pb age of 347 ± 3 Ma for the Hongshishan ophiolitic gabbros and suggested a “Red-sea”-type oceanic basin based on their geochemical result (section 2.2.8). It is inferred that the Jijitaizi-Xiaohuangshan ophiolite mélange probably connects the Tonghuashan-Gulugou-Akeyazi-Atbashi ophiolite belt in the southwestern Tianshan, forming a continuous South Tianshan ophiolite belt which was laterally displaced by the sinistral Xingxingxia Fault (Han and Zhao., 2018). In addition, the Hongshishan ophiolite belt is regarded as eastern extension of the North Tianshan ophiolite belt (Wang et al., 2014c). As for the Liuyuan Unit, a current debate exists on the tectonic setting of the exposed mafic rocks. Some suggest a post-orogenic extension or a mantle plume setting in the Early Permian (Li et al., 2013; Su et al., 2013), while others suggest that it might represent an Early Permian “Liuyuan Ocean” and thus, the final closure of the middle segment of the Paleo-Asian Ocean possibly occurred in the late Permian or even in the Triassic (Ao et al., 2010; Guo et al., 2012; Mao et al., 2012; see section 2.2.2). Nevertheless, some recent data favor an Early Permian rifting setting (Chen et al., 2016; Wang et al., 2017; Niu et al., 2018).

Roughly E-W large-scale ductile shear zones are also developed in the Beishan Belt. Such as: the North Beishan (Hongshishan) dextral shear zone, the Xingxingxia-Shibanjing dextral shear zone, and the Gubaoquan-Hongliuyuan shear zone (Cai et al., 2012). Actually, the Beishan shear zones are widespread and highly developed but have attracted relatively little attention so far. Further works on these ductile shear zones are required in order to better understand the tectonic evolution of the Beishan Belt.

Although paleomagnetic data are indispensable for reconstructing the paleogeography and unveiling orogenic processes, so far, no late Paleozoic to early Mesozoic paleomagnetic data have been reported in the Beishan Belt. In contrast, a first Early Permian paleomagnetic pole for the Dunhuang Block has been obtained, suggesting that an ocean still existed between the Beishan and Dunhuang in the Early Permian and that the final closure of the Paleo-Asian Ocean in the Beishan occurred in the Late Permian (Xu et al., 2019).

1.1.3 The Mongolia Belt

The final closure of the eastern segment of the Paleo-Asian Ocean is proposed mainly in the Late Devonian (Xu et al., 2013; Zhao et al., 2013) or the Late Permian to Early Triassic (Xiao et al., 2003; Eizenhöfer et al., 2018; Zhang et al., 2018) based on sedimentological, paleontological, paleomagnetic, geochemical, and structural studies. Detailed introduction of the Mongolia Belt is beyond the scope of this thesis. Here, we present only some recently published paleomagnetic works, which are necessary in the discussion part (Chapter 8).

Zhang et al. (2018) performed paleomagnetic works on both North China Craton and Songliao-Xilinhhot Block, getting a 280 Ma paleomagnetic pole for the Songliao-Xilinhhot Block with a low paleolatitude of $4.9^{\circ} \pm 7.5^{\circ}$. Accordingly, they proposed that the Songliao-Xilinhhot Block probably moved along a left-lateral transform boundary along the present-day northern margin of the North China Craton and that the Paleo-Asian Ocean closed until the Late Permian. However, Zhao et al. (2020) suggest that the magnetic declination of samples in Zhang et al. (2018) is nearly perpendicular with respect to that of previous studies and that the use of magnetic inclination shallowing factor $f = 0.6$ for the previous paleomagnetic data of the Mongolia Block is inappropriate. In presenting their new Early Permian data from volcanic rocks in northeastern Mongolia, Zhao et al. (2020) raised the possibility that the Paleo-Asian Ocean between the Mongolia and North China was closed at least in the Early Permian

and that significant strike-slip-related rotations occurred within and along the margins of the unified Mongolia Block and North China Craton due to posterior orogenic events. Recently, Ren et al. (2020) reported a new paleomagnetic study and indicated that the eastern Paleo-Asian Ocean experienced a scissor-like closure from west to east and the final closure happened in Late Permian to Early Triassic.

1.2 Major pending scientific issues

Although abundant studies have been performed and substantial progress has been made on the late Paleozoic to early Mesozoic evolution of the southern CAOB, some scientific problems still persist that need be resolved for the sake of a deeper understanding of this evolution process.

1.2.1 Revealing the kinematics and geological significance of the large-scale strike-slip shear zones

Several large-scale strike-slip shear zones extend roughly west-east in the southern CAOB, and have received much attention in last decades. Many structural and thermochronological studies found that these shear zones generally have dextral kinematics, and late Carboniferous to latest Permian $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages in the west part (Laurent-Charvet et al., 2002, 2003; Wang et al., 2008b; de Jong et al., 2009; He et al., 2018a; details in the section 2.1.2) and Triassic ages in the east part (Zhao et al., 2015).

According to paleomagnetic data, the magnitude of strike-slip displacements along the major dextral shear zones in the Tianshan Belt reach several hundreds of kilometers (Wang et al., 2007; Choulet et al., 2011). These dextral shear zones coupled with sinistral Erqis shear zone indicate an eastward extrusion of the Yili, Junggar and North Tianshan units relative to the Siberia and Tarim cratons (Wang et al., 2007, 2008a; Wang et al., 2008b; Charvet et al., 2011; Li et al., 2017), which probably resulted from intraplate oblique convergence (Wang et al., 2009; Charvet et al., 2011; Choulet et al.,

2012). Large-scale strike-slip shear zones also developed in the Beishan Belt and Mongolia Belt (Cai et al., 2012; Zhao et al., 2015) but lack of study. Recent paleomagnetic studies in the Mongolia Belt suggest that significant post-Early Permian vertical-axis strike-slip-related rotations also occurred within and along the margins of the unified Mongolia Block and North China Craton (Zhang et al., 2018; Zhao et al., 2020), but their senses and magnitudes are not sufficiently constrained.

Many scholars regard those strike-slip displacements as the intracontinental adjustment after the closure of Paleo-Asian Ocean (e.g. Han and Zhao, 2018). In fact, strike-slip faults in the southern CAOB are well-developed and can form a west-east fault system with basically similar ages and kinematics. Such a fault system hints that the strike-slip displacements are the main intracontinental tectonic pattern after the amalgamation and must have played an important role in building the final configuration of the southern CAOB, forming accompanied late Paleozoic - early Mesozoic magmatism and mineral deposits (e.g., Pirajno, 2010), and deciphering paleogeodynamics in this region. Moreover, these shear zones provide a natural laboratory to study stress partitioning and strain localization in regional-scale transpressional and transtensional deformation zones. Therefore, more investigations are requisite in revealing the kinematics and geological significance of the large-scale strike-slip shear zones in the southern CAOB.

1.2.2 Understanding the tectonic transition from collision/accretion to intracontinental tectonics

The Paleozoic collision/accretion processes of the CAOB have been extensively studied (Windley et al. 2007; Cai et al., 2011; Wilhem et al., 2012; Zhang et al., 2013; Eizenhöfer et al., 2014; Safonova and Santosh, 2014; Xiao et al., 2015), and the ensuing large-scale strike-slip displacements along the major faults have also been documented by many structural, thermochronological, and paleomagnetic data (Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2008b; de Jong et al., 2009; Choulet et al., 2011; Cai

et al., 2012; Yi et al., 2015). However, how the tectonic transition from collision/accretion to intracontinental tectonics exactly operated remains unclear. The timing, pattern, and geodynamic process concerning this transition are the focus issues in academic community. Different final closure times of the Paleo-Asian Ocean have been proposed in and between its different segments (Xiao et al., 2003; Ao et al., 2010; Guo et al., 2012; Mao et al., 2012, 2014; Xu et al., 2013; Zhao et al., 2013, 2020; Chen et al., 2016; Eizenhöfer et al., 2018; Han and Zhao, 2018; Niu et al., 2018; Zhang et al., 2018). Several hypotheses, including oblique convergence (Wang et al., 2009; Charvet et al., 2011; Choulet et al., 2012), slab break-off (Yuan et al., 2010; Ma et al., 2015), and extrusion tectonics (Wang et al., 2007, 2008a) have been proposed to interpret the end Paleozoic to early Mesozoic tectonic transition. Furthermore, different tectonic settings were also suggested for the late Paleozoic tectonic and magmatic environment, such as ridge subduction or subduction (Geng et al., 2009; Ao et al., 2010; Mao et al., 2012), post-orogenic extension/rifting (Li et al., 2013; Chen et al., 2016) and super mantle plume-related large igneous province (Xia et al., 2008, 2012; Su et al., 2011a, b). The consensus is difficult due to limited data and research approaches employed so far, mainly geochronology and geochemistry of igneous rocks - both sometimes leading to ambiguous tectonic interpretations (Zhu et al., 2018, 2019). Therefore, more constraints from different perspectives are greatly needed.

Kinematics along the major boundaries between microcontinental blocks and/or terranes in the southern CAOB during late Paleozoic to early Mesozoic time is the key to decipher the geodynamic evolution and to understand mechanisms of transition from subduction/accretion to intracontinental tectonics. Plate boundaries in orogenic belts usually experience long and multi-stage deformation, and geodynamic reconstruction using discontinuous and complex geological records often remain ambiguous and controversial. Paleomagnetism is a workable method to study these processes, as it provides quantitative, large-scale constraints on the kinematics of blocks, including paleolatitudinal translation and vertical-axis rotation (Butler, 1992; Tauxe et al., 2010).

Furthermore, through comparisons of paleomagnetic poles of different tectonic units in the southern CAOB, the kinematic evolution, in terms of velocity, orientation, and pattern, among these tectonic units during a certain geological period could be revealed, providing crucial basis for understanding aforementioned tectonic transition.

1.2.3 Achieving well-constrained late Paleozoic to early Mesozoic paleogeographic reconstructions

Nowadays, it is widely accepted that the youngest supercontinent Pangea formed following the assembly of Gondwana and Laurasia. The main configuration of Pangea is framed through voluminous evidence from robust paleomagnetic, geological, and paleontological data. However, reconstructing the East Asian blocks in Pangea is still in controversy (Zhao et al., 2018). The main reason is that the geological domains/blocks in East Asia are complicated, especially in the CAOB which is composed of numerous and variously deformed microcontinents, island arcs, seamounts, and accretionary wedges (Windley et al., 2007; Wilhem et al., 2012; Han and Zhao, 2018). Due to this complexity, only few reliable paleomagnetic data have been obtained from the microcontinents (except the Tarim Block) in the CAOB (Wang et al., 2007; Choulet et al., 2011, 2013; Zhang et al., 2018), which could provide the paleogeographic framework (Butler, 1992; Tauxe et al., 2010). As a result, no robust paleogeographic reconstruction for the CAOB has been reached.

However, the paleogeography is essential to understand the origin and diversity of life, ancient climate, and the distribution of tectonic belts and mineral resources (Huang et al., 2018). More specifically, the relative paleogeographic positions of different tectonic units in the southern CAOB is an essential foundation for the restoration and reconstruction of the paleo-ocean-continent framework, evolution of the oceanic basin, continental drift, and related orogenic processes. Recent progress in the Himalayan orogenic belt is a good analogy (van Hinsbergen et al., 2012; Yang et al., 2015; Ma et al., 2017).

Therefore, more late Paleozoic to early Mesozoic paleomagnetic works on the microcontinents in the southern CAO (i.e. the Yili, Turpan-Hami, Baluntai, Junggar blocks, and terranes in the Beishan Belt) are required to achieve a robust and consistent paleogeographic reconstruction at the scale of the southern CAO during this geological period.

1.3 Research purpose

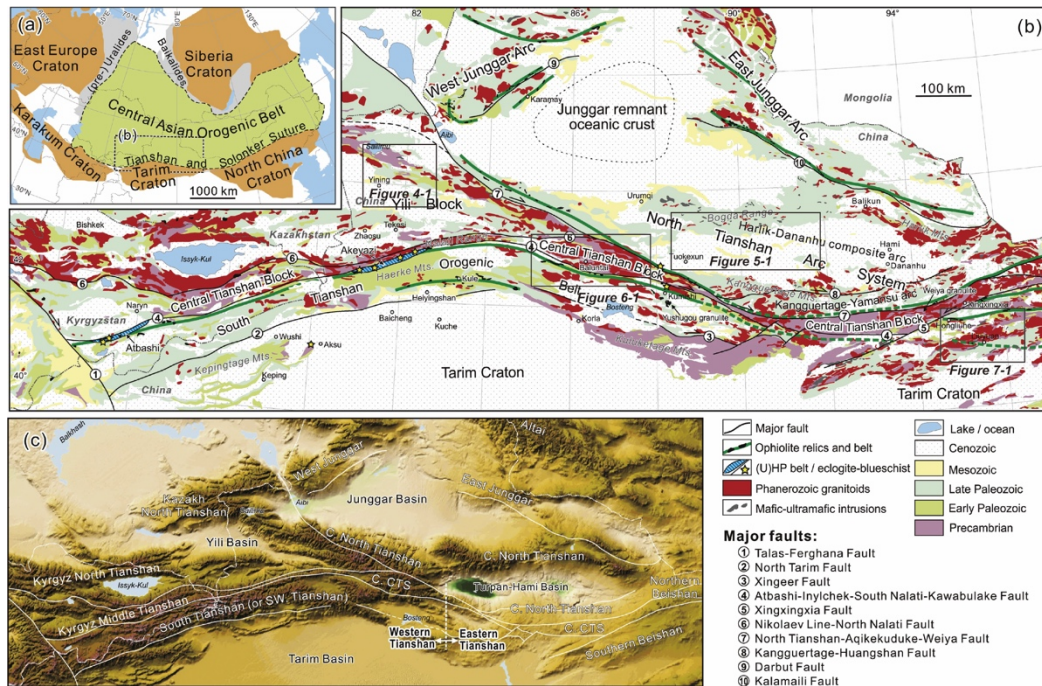


Figure 1-2. (a) Tectonic sketch of the CAO and its surrounding regions. (b) Geological map of the southern CAO covering the Tianshan and Beishan belts. (c) Topographic map of the Tianshan and Beishan belts (from Han and Zhao (2018)). Our paleomagnetic sampling locations are marked on the Figure 1-2b.

The late Paleozoic to early Mesozoic paleogeography, tectonic transition from collision/accretion to intracontinental tectonics, and large-scale strike-slip displacements along major faults in the southern CAO are of great importance in understanding the evolution history of the CAO. Paleomagnetism is one of the powerful techniques to resolve these scientific problems, while only few paleomagnetic data reported in the southernmost CAO so far. Therefore, paleomagnetic samples were

collected from the Yili, Turpan-Hami, and Baluntai blocks in the Tianshan Belt, and Shuangyingshan terrane in the Beishan Belt, covering Neoproterozoic and Early Permian mafic dikes, Early Permian volcanic and sedimentary rocks, and Triassic sedimentary rocks, trying to obtain reliable paleomagnetic data for these microcontinents and provide more constraints on aforementioned scientific problems (Figure 1-2).

1.4 Workload of the study

Detailed workload can be referred to Table 1-1.

1.5 Major innovations

The major innovations of this study are listed below:

1. This study provides several high-quality late Paleozoic paleomagnetic poles for the microcontinents/terrane in the southern CAOB, which are of great importance for better understanding the tectonic and geographic evolution of the southern CAOB but have been lacking for a long time.

2. New techniques, such as Elongation/Inclination correction (Tauxe & Kent, 2004), scanning electronic microscopic observation coupled with energy-disperse spectroscope, multi-minerals $^{40}\text{Ar}/^{39}\text{Ar}$ dating, and paleogeographic reconstruction software Gplates, were used in this study, further developing traditional paleomagnetic study process.

3. In this study, paleomagnetism was used to constraint late Paleozoic paleogeography and tectonic transition of the southern CAOB. Unlike traditional study methods, such as geochemistry, structural analyses, sedimentology, and paleontology, paleomagnetism offers us a regional view and a quantitative way to test and incorporate previous data, and then reach a deeper understanding of abovementioned pending scientific problems in the southern CAOB.

Table 1-1. The list of workload of this study

<i>Content</i>	<i>Unit</i>	<i>Number</i>	<i>Notes</i>
Field observation	zone	2	The Tianshan and Beishan belts
	day	52	30 days in the Tianshan Belt and 22 days in the Beishan Belt
	point	351	Field observation points and paleomagnetic sampling points
Thin section observation	sample	18	Observe mineral composition, degree of deformation and metamorphism; primarily determine major magnetic carries
Thermomagnetic experiment	sample	42	Study major magnetic carriers and their grain sizes
Hysteresis loop	sample	63	
Isothermal remanent magnetization acquisition curve	sample	38	
Anisotropy of magnetic susceptibility	specimen	596	Specimens from Yili, Turpan-Hami, and Baluntai blocks and Beishan Belt, including Neoproterozoic and Early Permian mafic dikes, late Carboniferous migmatites, and Early Permian and Triassic volcanic and sedimentary rocks
Scanning electronic microscope observation	sample	12	Samples from Yili Block and Beishan Belt were observed to see if they underwent later thermal or chemical influence
Energy disperse spectroscopy	point	206	To roughly determine magnetic minerals
U-Pb zircon dating	sample	5	Constrain the formation age of studied paleomagnetic samples
$^{40}\text{Ar}/^{39}\text{Ar}$ dating	sample	19	Determine the remanence age of studied samples, and the cooling ages of ductile shear zones in the Beishan and Xingxingxia areas
Demagnetization	specimens	>800	Meaningful characteristic remanent magnetizations were isolated from two-third specimens

Chapter 2. Geological Background

2.1 Tianshan Orogenic Belt

The Tianshan belt is located in the southwestern part of the CAOB, extending ~2500 km from northwest China, across Kazakhstan, Kyrgyzstan, and Tajikistan, to Uzbekistan (Figures 2-1a-b). Its eastern segment, the Chinese Tianshan Belt, is tectonically subdivided, from north to south, into North Tianshan, Yili Block, Central Tianshan, and South Tianshan by large-scale dextral shear zones which reworked the ancient suture zones (Figure 2-1b; Gao et al., 1998; Wang et al., 2011; Charvet et al., 2011; Zhong et al., 2015).

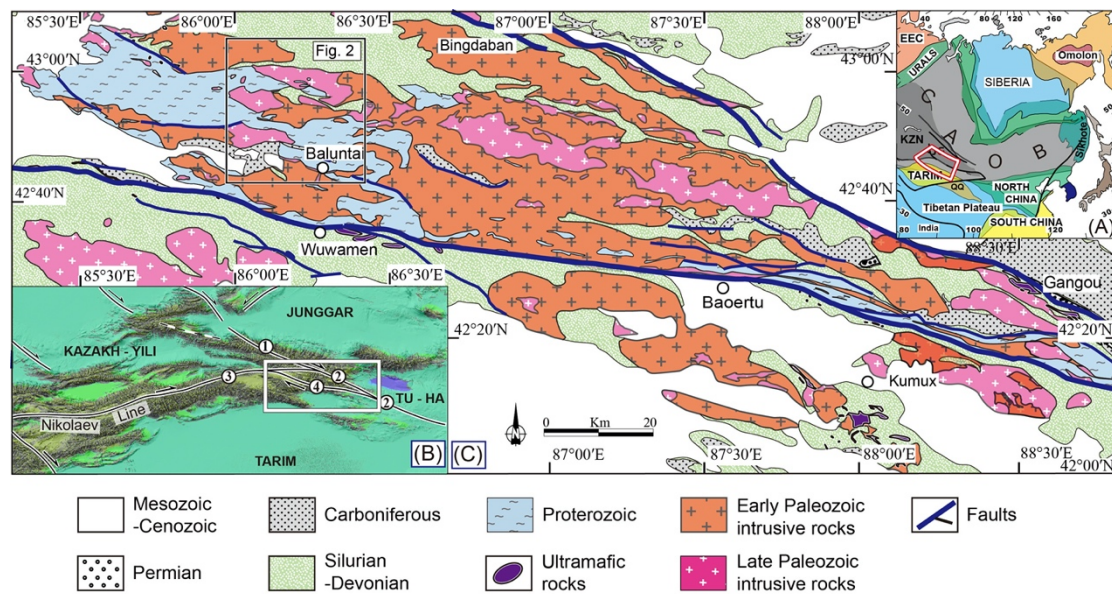


Figure 2-1. (a) Simplified tectonic division of the eastern Asia (after Jahn, 2004). EEC = East European Craton; KZN = Kazakhstan; Q = Qaidam. (b) Tectonic division of the Tianshan Orogenic Belt. Major faults: 1. North Tianshan Fault = NTF; 2. Middle Tianshan Shear Zone = MTSZ; 3. Nalati Fault = NF; and 4. Baluntai Fault = BF. (c) Geological map of the Baluntai Block showing our studied area (Zhong et al., 2015).

2.1.1 Tectonic units of the Chinese Tianshan Belt

The North Tianshan is separated from the Yili Block and the Central Tianshan by

the North Tianshan Fault (NTF) and the Middle Tianshan Shear Zone (MTSZ), respectively. It is an accretionary complex, containing mainly Late Devonian to Carboniferous sedimentary and volcanoclastic rocks, formed by southward subduction of the North Tianshan-Junggar Ocean probably prior to 316 Ma (Wang et al., 2006; Han et al., 2010).

The Yili Block, with a wedge shape, is the eastern part of the Kazakhstan microcontinent and is bounded by the NTF to the north and the Nalati Fault (NF) to the south (Charvet et al., 2007; Gao et al., 2017). It comprises Precambrian metamorphic basement (Meso- to Neoproterozoic granitic gneisses and metasedimentary rocks; Wang et al., 2014a, b; Zhu et al., 2018), Late Neoproterozoic-Early Paleozoic sedimentary cover, Paleozoic magmatic arc sequences (Zhong et al., 2017), and Permian post-orogenic magmatic rocks (Wang et al., 2009; XBGMR, 1993). Immediately to the south border of the Yili Block, a high-pressure metamorphic complex (i.e. Akeyazi metamorphic belt) is exposed, containing micaschists, greenschists, quartzites, paragneisses, blueschist and eclogite lenses (Gao and Klemd, 2003). The peak metamorphism (blueschist- and eclogite-facies) happened from 345 to 320 Ma, following by a retrogression from 330 to 305 Ma (Klemd et al., 2005, 2011; Li et al., 2011; Zhang et al., 2018). The MORB- or OIB-like geochemical signature of the mafic protoliths of this complex (Gao and Klemd, 2003) is consistent with that of the 600-516 Ma ophiolites which crop out along the Akeyazi metamorphic belt (Yang et al., 2005; Qian et al., 2009), indicating that they probably represent the Paleo-Tianshan Ocean suture zone (Gao et al., 1998, 2009; Han et al., 2011) which likely extends eastwards to the Nalati-Mishigou-Gangou ophiolite mélange zone (Charvet et al., 2007, 2011; Lin et al., 2009; Wang et al., 2018).

The Central Tianshan is a continental block sandwiched between North Tianshan-Yili Block and South Tianshan units (Shu et al., 2002; Dong et al., 2011; Jiang et al., 2013), primarily referring to the Baluntai area (Figure 2-1c) and its westward extension to the northern parts of the Hark Shan and Erbin Shan (Charvet et al., 2007,

2011; Wang et al., 2011, 2018). It is bounded by the MTSZ and NF to the north, and the Baluntai Fault (BF) to the south (Shu et al., 2002, 2004; Jiang et al., 2015; He et al., 2018a). The Central Tianshan is composed of Proterozoic metamorphic and crystalline basement, Ordovician to Early Devonian metamorphosed and ductilely deformed arc sequences, Late Devonian to Carboniferous syn- or postcollisional undeformed granites, and sedimentary cover. The Proterozoic basement mainly comprises gneisses, amphibolites, migmatites, marbles, and mica schists, mainly exposed in the eastern segment of the Central Tianshan (e.g. Bingdaban, Baluntai regions, XBGMR, 1993). These rocks have been dated at ca. 630 and 926 Ma (zircon U-Pb) in the Bingdaban area (gneissic granites, Chen et al., 2009, 2012), ca. 700 Ma in Baluntai area (gneissic granite, Yang et al., 2006), and ca. 1142 and 1455 Ma in the Gangou-Kumux area (gneisses, Shi et al., 2010). Moreover, Li et al. (2009) reported ca. 1200 Ma, 1530Ma, and 1800 Ma detrital zircon age peaks for amphibolites and metasedimentary rocks collected from the Xingxingxia Group in the Baluntai area, indicating the possible existence of a Meso- Paleoproterozoic basement. The high-metamorphic and crystalline basement is either covered by Ordovician to Silurian sandstones and interbedded volcanoclastic rocks, which locally underwent greenschist face metamorphism, or crosscut by Silurian to Early Devonian ductilely deformed plutonic rocks (Shu et al., 2011; Ma et al., 2012a, b; Wang et al., 2018). Previous geochemical and geochronological studies suggest that these early Paleozoic igneous rocks were formed in a continental arc setting (Yang et al., 2006; Ma et al., 2014; Zhong et al., 2015), which likely resulted from the southward subduction of the Paleotianshan Ocean (Wang et al., 2008, 2011; Alexeiev et al., 2015; Loury et al., 2015). Widespread undeformed Late Devonian to Carboniferous intrusive rocks, with syn- or postcollisional geochemical properties, crosscut the aforementioned lithotectonic units (Shi et al., 2007). Furthermore, the metamorphosed and/or deformed Paleozoic and Proterozoic rocks are separated from the Lower Carboniferous molasse of Maanqiao Formation by an angular unconformity (XBGMR, 1993; Shu et al., 2004). The

sedimentary sequences grade upward into shallow marine sandstones and carbonates, and are not ductilely deformed nor metamorphosed (Carroll et al., 1995).

The South Tianshan refers to the area bordered by the NF and the BF to the north and the South Tianshan ophiolitic mélange zones to the south, covering the Hark Shan, Erbin Shan, Huola Shan and Keziletage. Paleozoic marine sedimentary sequences and deformed Ordovician-Early Devonian intermediate-felsic volcanic rocks resembling Andean-type continental arc ones occur in this region (Zhong et al., 2019). Ophiolite mélanges with ages of 439-310 Ma (zircon U-Pb) (Long et al., 2006; Han et al., 2011; Yang et al., 2011; Wang et al., 2011, 2012, 2018) are discontinuously exposed in the South Tianshan, which likely resulted from the diachronous opening and closing of back-arc basins in this unit (Wang et al., 2018; Zhong et al., 2019).

2.1.2 Large-scale shear zones

As the tectonic boundary between the North Tianshan and Central Tianshan, the MTSZ extends eastwards from Xiaergou-Gangou to Weiya areas (Figure 2-1b; Shu et al., 1999; Laurent-Charvet et al., 2002; He et al., 2020). Mylonitized rocks in this shear zone generally show vertical mylonitic foliation and sub-horizontal stretching lineation with consistent asymmetric fabrics indicating a dextral kinematic feature (Shu et al., 1999; Laurent-Charvet et al., 2002, 2003; Charvet et al., 2007; Cai et al., 2012). The NTF separating the North Tianshan and Yili Block, being regarded as west extension of the MTSZ, also exhibits pervasive dextral kinematic features (Zhou et al., 2001; Wang et al., 2006, 2009; de Jong et al., 2009). Previous thermochronologic studies indicate that the ductile shearing in the MTSZ and the NTF initiated in the latest Carboniferous (~312-310 Ma; Xu et al., 2011; He et al., 2020) and lasted until the Middle Triassic (~245-242 Ma; Shu et al., 2002; Laurent-Charvet et al., 2003; de Jong et al., 2009; Yang et al., 2009; Cai et al., 2012).

The NF reworked the Paleo-Tianshan Ocean suture zone between the Yili Block and the South Tianshan (e.g., Gao et al., 2009; Qian et al., 2009; Long et al., 2011; Zhong

et al., 2017), and it connects the NTF near the Bingdaban area. Structural studies from this fault (Akeyazi high-pressure belt, northern Bayanbulak, Laerdun Daban) indicate dextral kinematics (Li and Liu, 1997; Wang et al., 2010; Lin et al., 2009; Zhong et al., 2019). Mica $^{40}\text{Ar}/^{39}\text{Ar}$ dating results from mylonitic samples suggest this dextral shear occurred at ~285-252 Ma (Zhou et al., 2001; de Jong et al., 2009). Recently, Zhong et al. (2019) report 307-255 Ma metamorphic zircons from mylonitic sandstones in the southern Nalati Range, indicating the NF probably initiated in the latest Carboniferous.

Extensive/detailed kinematic studies reveal that the BF has dextral kinematics (Laurent-Charvet et al., 2003; Wang et al., 2009, 2018; Tang et al., 2011; Xu et al., 2011; Cai et al., 2012). Mica and hornblend $^{40}\text{Ar}/^{39}\text{Ar}$ ages of granitic mylonites from the Kumishi, Aqi Shan, and Kawabulak areas show consistent shearing age of ~298-252 Ma (Laurent-Charvet et al., 2003; Wang et al., 2008b; Xu et al., 2011; Cai et al., 2012). Moreover, Yang et al. (2007) reported ~311-288 Ma biotite and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ ages for mylonitic rocks, whose grain-shape-fabrics exhibit a symmetrical morphology, collected from the south Baluntai and north Kumishi areas.

2.2 Beishan Orogenic Belt

Linking with the Tianshan Orogenic Belt to the west and the Mongolia Orogenic Belt to the east, the Beishan Orogenic Belt represents the central part of the southernmost CAO (Figure 2-2). The Xingxingxia sinistral strike-slip ductile fault which initially formed at around 240-235 Ma (Wang et al., 2010b), interrupts the tectonic continuity with the Chinese Eastern Tianshan to the west. To the east, the Beishan Orogenic Belt is covered by the Badain Jaran Desert and limited by the Ruoshui River (Zuo et al., 1990; Liu and Wang, 1995; Xiao et al., 2010). Moreover, the Beishan Orogenic Belt is framed by the Dunhuang Block in the south (Shulehe Fault as probable boundary; Zuo et al., 1990) and the Mongolia Orogenic Belt in the north (He et al., 2018b).

The Beishan Orogenic Belt consists of several terranes, namely, from south to north, the Shibanshan, Shuangyingshan, Mazongshan, Hanshan, and Queershan terranes,

which are separated by the Liuyuan Unit, and Hongliuhe-Xichangjing, Jijitaizi-Xiaohuangshan, and Hongshishan ophiolitic mélanges (Figure 2-2; Xiao et al., 2010; He et al., 2018b).

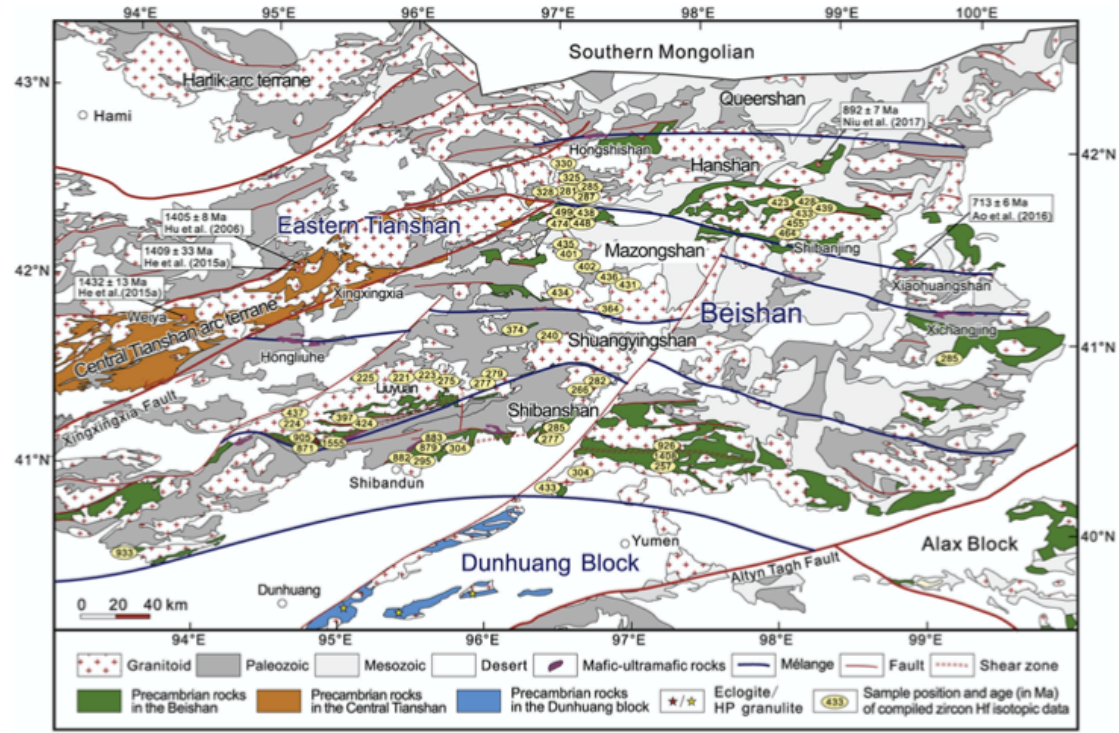


Figure 2-2. Simplified geological map of the Beishan orogen and the adjacent Dunhuang block and Chinese Eastern Tianshan orogen, showing major terranes of the Beishan orogen and the mélanges separating the terranes (He et al., 2018b).

Note: if there is no specific description, following ages stand for in-situ zircon ages.

2.2.1 Shibanshan terrane

The Shibanshan terrane possibly have a Precambrian basement as indicated by several studies. Jiang et al. (2013) first obtained high-quality Neoproterozoic ages of ~880 Ma for Shibandun amphibolites and Baidunzi gneisses; Neoproterozoic ages were also reported for granodiorites in Yadan (933 ± 5 Ma; Yuan et al., 2015), mylonitic granites (926 ± 10 Ma; He et al., 2018b) and garnet schists in Jiujing (age of zircon rims: 900 ± 3 Ma; Zong et al., 2017), and granitic gneisses in Shibandun (879 - 883 Ma; He et al., 2018b). Furthermore, recently, many Mesoproterozoic rocks were found in this

terrane: He et al. (2015) gained ~1400 Ma ages for biotite-plagioclase gneisses and granodiorites in Jiujing; Zong et al. (2017) acquired ~1400 Ma age for zircon cores collected from garnet schists in Jiujing. Many magmatic and metamorphic rocks of ages varying from 1395 to 1450 Ma were also reported in the Jiujing, Qiaowan, and Shibandun areas (Yuan et al., 2019). In addition, detrital zircons from sedimentary rocks in Chuanshanxun show ages ranging from 269 to 2486 Ma with Precambrian peaks (Zheng et al., 2018a).

Ordovician-Silurian medium- to high-grade metamorphic rocks, mainly consisting of schists, amphibolites, gneisses, and migmatitic gneisses, are widespread in this terrane (He et al., 2018b). However, their early Paleozoic ages were poorly constrained by fossils (*Tropidiscus sp.* and *Maclurites sp.*) in crystalline limestones. Locally, these metamorphic rocks experienced extensive mylonitization (Cai et al., 2012) and were alternatively interpreted as Archean basement (Beishan Complex) according to limited isotopic ages (e.g. Sang et al., 1997). Recently, Devonian to Permian ages were also proposed for these metamorphic rocks (Song et al., 2016; Zheng et al., 2018a). Only few early Paleozoic magmatic rocks from this terrane were reported in the literature. Basalt with age at 420 Ma was found from the lower part of Sangejing Formation in Dundunshan (Li et al., 2011); based on the unconformity just above the Sangejing Formation, the authors indicated that the Beishan Paleozoic oceanic basin was closing during the end of Silurian. He et al. (2014) obtained ages of 435 and 433 Ma for Jinta granite and Qiaowan mylonitic granite, respectively; combining the coeval high-pressure granulite in the Dunhuang Block, they suggested that the early Paleozoic magmatisms in the Shibandun terrane probably resulted from the collision and continued convergence between the Shibanshan terrane and the Dunhuang Block; thus, this terrane was regarded as an early Paleozoic continental arc.

Late Paleozoic rocks are also widespread in this terrane. Li et al. (2011) obtained zircon U-Pb age at 367 ± 10 Ma for andesites in Dundunshan, and regarded volcanic rocks of the Dundunshan Group as the results of post-orogenic extension. In contrast, an

arc geochemical signature was revealed for the Late Devonian (~370 Ma) volcanic rocks in Dundunshan by Guo et al. (2014), and a retreat of subduction zone was put forward to interpret the geochemical signatures of the magmatic rocks. Permian granites, granitic gneisses, doleritic dikes, amphibolites, and meta-graywacke were studied in several recent works (304 - 257 Ma; Zhang et al., 2010, 2011; Feng et al., 2012; Jiang et al., 2013; He et al., 2015; Song et al., 2016; Zheng et al., 2018a; Yuan et al., 2019). Most scholars consider that these rocks were formed in a post-collisional extension to rifting tectonic setting, while some evoke a continental arc setting (Ao et al., 2010; Xiao et al., 2010; Guo et al., 2012; Mao et al., 2012).

2.2.2 Liuyuan Unit

To the north, the Liuyuan Unit contains Lower Permian strata, Zhesi Group, which can be further divided into two subunits. The Lower Zhesi Group consists of thick packages of conglomerates interbedded with both sandstones and siltstones at the lower part, while its middle and upper parts are characterized by alternating thin beds of fine and medium grained mudstones, siltstones, and sandstones (Chen et al., 2016). The Lower Zhesi Group is generally fining-upwards. The Upper Zhesi Group contains thick packages of basalts which are locally interbedded with mudstones, tuffaceous sedimentary rocks, and fossiliferous limestones (*Camarophoria cf. mutabilis* Tschern., *Spiriferella* sp., *Rhynchopora* sp., *Waagenoconcha* sp., and *Muirwoodia cf. mammata* (Keys.)) with Kungurian and Roadian ages (Chen et al., 2016). Permian mafic-ultramafic rocks, including gabbros, diabases and ultramafic rocks are also abundant in and close to the Liuyuan Unit (see Shuangyingshan terrane). They are stocks or minor bodies intruding Ordovician to Silurian and Permian strata and generally along major faults.

Some scholars proposed that the Liuyuan Unit is an ophiolitic mélange mainly based on geochronological and geochemical studies on the Permian mafic-ultramafic rocks (Mao et al., 2012; Pirajno et al., 2008; Ao et al., 2010), and Cleven et al. (2015) considered these intrusions as dismembered fragments of the Hongliuhe Ophiolite. In

contrast, other scholars argued that the Liuyuan Unit is an outcome of late Paleozoic intraplate extension according to: (1) geochemical studies on the Permian granitic and mafic-ultramafic intrusions (Su et al., 2011b, 2013; Zhang et al., 2012; Li et al., 2013; Xue et al., 2018; Yuan et al., 2018); (2) extensive and detailed work on the Early Permian volcano-sedimentary sequence in the Liuyuan area (Chen et al., 2016) and study on the Upper Paleozoic megasequences in the Beishan Region (Niu et al., 2018); and (3) the first Early Permian paleomagnetic pole for the Beishan Belt (Zhu et al., 2020).

2.2.3 Shuangyingshan terrane

Precambrian basement probably also exists in the Shuangyingshan terrane. Medium- to high-grade metamorphic rocks crop out along a 10 km-wide belt in the Gubaoquan area, mainly including migmatites, quartz schists, gneisses, and enclosing retrogressed Ordovician eclogite boudins (Liu et al., 2011; Qu et al., 2011; Saktura et al., 2017; Soldner et al., 2020). Mesoproterozoic inherited zircons (~1400 Ma) were found from the Gubaoquan augen gneisses (Yuan et al., 2015) and a Mesoproterozoic upper intercept age of the Gubaoquan paragneisses was also obtained (Liu et al., 2011). In situ zircon U-Pb dating on a granitic orthogneiss from Gubaoquan yielded a metamorphic age of ~1450 Ma and a magmatic age of ~1550 Ma (He et al., 2018b). In addition, Ao et al. (2016) showed a maximum deposit age of 1229 Ma for a sandstone collected from the south of Xichangjing. Neoproterozoic rocks were well studied in Gubaoquan, giving high-quality ages at 871-905 Ma (Ye et al., 2013; Liu et al., 2015; Yuan et al., 2015; He et al., 2018b). Moreover, Precambrian detrital zircon age populations were acquired from metasedimentary and sedimentary rocks of this terrane (Song et al., 2013b; Ao et al., 2016). Sinian to Silurian sedimentary strata in this terrane are mainly composed of limestones with subordinate sandstones and mudstones.

Early Paleozoic magmatism is developed in the Shuangyingshan terrane, especially during the Silurian. Rock complex from Mingshujing pluton consist of gabbro, gabbroic granodiorite, granodiorite, and granite. Their emplacement ages are at 449-436 Ma and

were indicated to represent a tectono-magmatism event related to oceanic slab subduction in the southern Beishan area (Li et al., 2015). Mao et al. (2012) reported an adakite with a crystallization age at 424 ± 4 Ma, which was probably formed by melting of a hot subducting oceanic slab. Two mineralized granitoid intrusions, Pantuoshan and Yingzuihongshan plutons, yield intrusion ages at 424-417 Ma and emplaced in an arc collision setting. Zhao et al. (2007) studied the granitoids in the Liuyuan area with emplacement ages from 436 to 397 Ma, raising the possibility that the Liuyuan granitoids emplaced in a post-collisional tectonic setting likely related to slab break-off. Li et al. (2009, 2011) performed geochronologic and geochemistry works on the Shuangfengshan granitic pluton and Huitongshan K-feldspar granitic pluton, providing 415 ± 3 Ma and 397 ± 3 Ma emplacement ages for the two plutons, respectively, suggesting both formed in a post-orogenic extensional or a late-stage syn-orogenic setting.

Numerous Middle and Late Devonian mafic and acidic rocks were found in this terrane. Xie et al. (2015) reported an age of 397 ± 4 Ma for the olivine hornblende gabbro norite, and suggested that the Beishan area was in a continental margin rift extensional environment then. An emplacement age of 385 ± 8 Ma was acquired for the Yingmaotuo pluton, and the granitoids in this area were suggested to have formed during intra plate extension (Wang et al., 2015). The gabbros, gabbroic dikes, amphibolitic gabbros, and diorites in the Heishan pluton have emplacement ages at 377-356 Ma (Xie et al., 2012; Yang et al., 2012). Yang et al. (2012) put forward that the Beishan area was always in a relatively intensive extensional setting during whole late Paleozoic, and Xie et al. (2012) suggested that the emplacement of Heishan pluton were triggered by upwelling of asthenosphere due to break-off of subduction slab. Similar ages were also obtained from several plutons, such as Datoushannan pluton (gabbro: 374 ± 3 Ma, Wang et al., 2015), Datoushan pluton (gabbro: 359 ± 6 Ma, Yan et al., 2012), and Guaishishan pluton (gabbro: 359 ± 4 Ma; Yang et al., 2016).

There was a cessation of magmatism during the Carboniferous in the

Shuangyingshan terrane. Few Carboniferous magmatic rocks were discovered (gabbro: 322 Ma, Su et al., 2011a). In contrast, the Permian magmatism is abundant and dominated by mafic-ultramafic complexes. Volcanic rocks (rhyolites: 279 ± 3 Ma; dacites: 265 Ma) were reported in Hongshishan (Su et al., 2011b) and rhyolites 285 ± 5 Ma old were found in Yushan. Baodanshan granite was dated at 285 ± 4 Ma, and was supposed to form during post-collisional stage (Zhang et al., 2012b). Post-collisional granitoids also exist in Dongdaqun and Wufengshan plutons (279-275 Ma, Li et al., 2013) and Hongheishan pluton (Yushan tungsten deposit, 281 ± 3 Ma). Mafic-ultramafic rocks were studied in Hongshihsan (281 ± 3 Ma, Ao et al., 2010; 286 Ma, Su et al., 2012b), Luodong (284 ± 2 Ma, Su et al., 2012a; 266 ± 3 , Xue et al., 2016), Poshi (284 ± 2 Ma, Qin et al., 2011), Podong (281 ± 2 Ma, Xue et al., 2016), Yinaoxia (281-265 Ma, Zheng et al., 2014; 282 Ma, Zhang et al., 2015), Gubaoquan (282 Ma; Zhang et al., 2015), Bijiaoshan (279 Ma, Qin et al., 2011), Poyi (271 ± 2 Ma, Xue et al., 2018), Xuanwoling (261 ± 2 Ma, Su et al., 2010), and northeast of Liuyuan (250 ± 9 Ma; Zhang et al., 2011). Most researchers suggested a post-orogenic extension or a mantle plume setting for the Beishan Belt during Permian, while some others invoked a subduction-related environment during this period, implying the existence of a “Liuyuan Ocean” (Ao et al., 2010; Guo et al., 2012; Mao et al., 2012). Triassic granitic rocks were also identified in the Shuangyingshan terrane: Middle Triassic granitic pluton (Dahuolu pluton, 240-238 Ma) likely linked to lithospheric delamination and asthenospheric upwelling, while Late Triassic plutons (Daquan, Baixianishan, Changliushui, and Huaniushan plutons, 225-217 Ma) were probably the result of crustal contamination and fractional crystallization for the granitic magma emplaced in a post-orogenic setting (Li et al., 2012).

2.2.4 Hongliuhe-Xichangjing ophiolitic mélange

The Hongliuhe-Xichangjing ophiolitic mélange consists mainly of Cambrian-Silurian pyroclastic rocks and clastic rocks, Cambrian-early Devonian ophiolitic

fragments, and Carboniferous-Permian clastic rocks (Liu and Wang et al., 1995; Nie et al., 2002; Xiao et al., 2010). The lower Paleozoic rocks are greenschist facies metamorphosed and deformed (Xiao et al., 2010). Ophiolitic fragments are distributed mainly in Hongliuhe, Yushishan, Niujuanzi, Huoshishan, Yueyashan, and Xichangjing areas. The Hongliuhe-Xichangjing ophiolitic mélange is well exposed and displays the most complete ophiolitic sequence considered by many as the main Paleozoic suture zone (Ren et al., 2001; Guo et al., 2006; Zhang et al., 2008; Cleven et al., 2015; Ao et al., 2016). However, other researchers propose that it represents a back-arc basin (Zuo et al., 1990; Zhang et al., 2012a, b, 2018).

In the Hongliuhe area, Guo et al. (2006) carried out amphibolite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dating on coarse-grained and deformed ophiolitic gabbros, undeformed monzonitic granite, and felsic mylonite. They obtained cooling ages of 496-462 Ma for the gabbros, cooling age of 412 ± 3 Ma for the undeformed granite, and ages of 341-339 Ma for later strike-slip movement and thermal event. Yu et al. (2006) conducted single-grain zircon U-Pb LA-ICP-MS dating on ophiolitic gabbro, acquiring an age of 426 ± 2 Ma. Zhang et al. (2008) and Cleven et al. (2015) performed high-quality geochronologic studies on Hongliuhe cumulated and pegmatitic gabbros, getting ages of 516 ± 7 Ma and 520 ± 6 Ma, respectively. Meanwhile, they found ages of 405 ± 5 Ma and 414 ± 4 Ma for undeformed granitoids, indicating the emplacement of these ophiolitic fragments was earlier than 414 Ma. Recently, Shi et al. (2018) reported a Cambrian age of 521 ± 4 Ma for ophiolitic gabbros in the Hongshishan area.

Ordovician-early Devonian ages were suggested for the ophiolitic fragments in the Niujuanzi and Huoshishan areas: a Rb-Sr isochron age of 463 ± 18 Ma was reported for Niujuanzi ophiolitic basalts by Ren et al. (2001); Ages of 447-431 Ma were obtained for Niujuanzi ophiolitic gabbros (Wu et al., 2012; Tian et al., 2014; Wang et al., 2018); Niujuanzi ophiolitic plagiogranites have ages of 449-430 Ma (Tian et al., 2014; Wang et al., 2018); Ages of 433 ± 3 Ma and 411 ± 4 Ma were reported for Niujuanzi ophiolitic diabases and Huoshishan gabbros, respectively (Tian et al., 2014; Wang et al., 2018). A

Carboniferous age of 354 ± 3 Ma was also proposed for the Niujuanzi ophiolitic gabbros (Wang et al., 2018). However, since there are two groups of ages, the meaning of this Carboniferous age remains to be refined.

In its eastern segments, i.e. Yueyashan and Xichangjing areas, Cambrian ages of 542-527 Ma were obtained for the ophiolitic plagiogranites and gabbros (Ao et al., 2012; Hou et al., 2012; Hu et al., 2015; Shi et al., 2018). Hu et al. (2015) also reported ages of 442 ± 2 Ma and 421 ± 15 Ma for the granodiorites and andesites in the Xieshan-Doqiyishan volcanic arc, suggesting that large-scale oceanic basin subduction occurred in the Silurian.

2.2.5 Mazongshan terrane

Abundant high-grade metamorphic rocks crop out in the Mazongshan terrane, consisting of orthogneisses, paragneisses, quartzites, marble, and mica schists. They were previously interpreted as Precambrian rocks by Zuo et al. (1990) and Liu and Wang (1995) according to geological arguments only. Recently, many geochronological studies on this terrane revealed that they only have Paleozoic protolith age with some inherited Precambrian zircons (Song et al., 2013a, b, 2015).

Deformed granites (granitic gneisses, mylonitic granites, foliated granites) were reported in Caohulehade (protolith ages: 526 ± 6 Ma and 410 ± 5 Ma, Song et al., 2013b; 420 ± 3 Ma, Song et al., 2015), Mazongshan (499-415 Ma, Song et al., 2013a), Hongliuxia (460-430 Ma, Yuan et al., 2018), Lebaquan (464-398 Ma, Song et al., 2015), Shuangjingzi (397 ± 3 Ma, Song et al., 2015). Underformed granitoids with ages from 435 to 356 Ma crop out in Baiqianta (435-431 Ma, Li et al., 2013), Baishantou (435-395 Ma, Wang et al., 2018; Zheng et al., 2018b), Bagelengtai (434 ± 6 Ma, Zheng et al., 2018b), Gongpoquan (402 Ma, Zheng et al., 2018b), Tongchangkou (426 ± 3 Ma and 408 ± 3 Ma, Zhang et al., 2018), Haergen (409 ± 3 Ma, Wang et al., 2018), Sangdaomingshui (365 ± 2 Ma, Zhang et al., 2018), Mazongshan (364 ± 4 Ma, Li et al., 2013), and Dongqiyi arc (356 ± 2 Ma, Zhang et al., 2012a).

Wang et al. (2018) compiled igneous rocks in the Mazongshan and Shuangyingshan terranes and proposed to divide the magmatism into three stages according to petrological, geochemical and isotopic data: (1) 470-423 Ma, subduction related magmatism; (2) 424-395 Ma, syn-/post-collisional magmatism; (3) 371-349 Ma, complex geodynamic setting connecting to southward subduction of the Xiaohuangshan back-arc ocean.

2.2.6 Jijitaizi-Xiaohuangshan ophiolitic mélange

Between Mazongshan and Hanshan terranes, the Jijitaizi-Xiaohuangshan ophiolitic mélange mainly occurring in the Jijitaizi, Shibanjing, and Xiaohuangshan areas, consisting mainly meta-ultramafic rocks, meta-basalts, mylonitic gabbros (Zuo et al., 1990; Xiao et al., 2010).

Previously, some scholars considered this mélange as the relic of a main early Paleozoic ocean, and suggested that the Hongliuhe-Xichangjing ophiolitic mélange represents a back-arc ocean (Zuo et al., 1990); while other scholars proposed a reversed model, i.e., the Hongliuhe-Xichangjing mélange represents a main ocean and Jijitaizi-Xiaohuangshan mélange is a relic of a back-arc ocean (He et al., 2002; Song et al., 2008; Yang et al., 2010).

Song et al. (2008) acquired an Ordovician Sm-Nd isochron age for the Xiaohuangshan ophiolitic gabbros, and proposed that the gabbros formed in a back-arc extension setting. A Cambrian age of 516 ± 8 Ma was also reported by Shi et al. (2018) for the Xiaohuangshan gabbros and indicated that the subduction of oceanic plate in the Xiaohuangshan area started in the Cambrian. However, several recent studies provided a Carboniferous age for this ophiolitic mélange: Li et al. (2012) obtained an age of 321 ± 4 Ma for the Jijitaizi ophiolitic gabbros; Zheng et al. (2013) got Carboniferous ages of 345 ± 14 Ma and 336 ± 4 Ma for the Xiaohuangshan ophiolitic gabbros and basalts, respectively; and Fu et al. (2020) also reported a Carboniferous age of 321 ± 6 Ma for the Shibanjing plagioclases. Based on its Carboniferous age, Li et al. (2012) indicated

that the Jijitaizi-Xiaohuangshan ophiolitic mélange is the result of Carboniferous-Permian post-collisional extension, while Ao et al. (2016) proposed the Xiaohuangshan Ocean was probably formed by slab rollback of the southward-subducted oceanic plate (Hongshishan Ocean?) closed in the Late Carboniferous.

2.2.7 Hanshan terrane

The Hanshan terrane consists widespread granitoids of ages from 464 to 281 Ma (Song et al., 2013a, b; Ao et al., 2016; He et al., 2018b; Yuan et al., 2018) and Ordovician to Carboniferous clastic rocks (Zuo et al., 1991; Ao et al., 2016; He et al., 2018b). Previous “Beishan complexes” were revealed to have Paleozoic protolith ages (Liu et al., 2010; Song et al., 2013a, b). However, Ao et al. (2016) found an orthogneiss with a protolith age of 713 ± 6 Ma. Furthermore, a Neoproterozoic gneissic xenolith was found in late Carboniferous granodiorite (Niu et al., 2017).

Yuan et al. (2018) suggested that the Hanshan terrane is probably a Precambrian microcontinent with Proterozoic basement based on plagioclase Sr and in-situ zircon U-Pb-Lu-Hf isotopic signatures of local Paleozoic granitoids. More importantly, they proposed that the early Paleozoic granitoids formed in a subduction-induced compressional setting, and the late Paleozoic ones emplaced in an intra-plate extensional setting, which is critical for understanding the architectural framework of the Beishan Belt. In contrast, Ao et al. (2016) proposed that the Hanshan terrane is an accretionary prism, accreting continuously from the Silurian to the Devonian on the northern margin of the Shuangyingshan-Mazongshan composite arc; the younging direction of the accretion is from south to north and the accretion continued until the Late Carboniferous.

2.2.8 Hongshishan ophiolitic mélange

The Hongshishan ophiolitic mélange is composed mainly by mafic-ultramafic ophiolitic rocks, limestones and cherts, marine sedimentary rocks, and mylonites in high-strain zones (Zuo et al., 1990; Xiao et al., 2010). The ophiolitic rocks are mainly

distributed in the Hongshihan, Baiheshan, and Pengboshan areas, and the largest one is Hongshishan ophiolite.

Some scholars suggested that the Hongshishan ophiolitic *mélange* is the relic of a late Paleozoic (Carboniferous) ocean, representing the suture between Kazakhstan and Tarim blocks (He et al., 2002; Gong et al., 2002). Some others considered this *mélange* as the geological record of the long-existing Paleo-Asian Ocean (early Paleozoic-Carboniferous), indicating the suture between Kazakhstan and Siberia blocks (Nie et al., 2002). The rest of scholars suggested that the Hongshishan ophiolitic *mélange* is not a suture; more probably, it records an embryonic ocean crust formed in a transition tectonic setting from continental rift to ocean in Carboniferous (Zuo et al., 1990; Yang et al., 2010). Wang et al. (2014) reported the first high-quantity age of 347 ± 3 Ma for the Hongshishan ophiolitic gabbros and suggested a “Red-sea”-type oceanic basin based on their geochemical result.

2.2.9 Queershan terrane

The Queershan terrane is located in the northernmost of the Beishan Belt and close to China-Mongolia border, and the degree of research in this terrane is limited. It is mainly composed of Ordovician to Early Permian clastic rocks, pyroclastic rocks, carbonates, and late Paleozoic volcanic rocks and granites (Xiao et al., 2010; Chen et al., 2015; He et al., 2018b).

This terrane is regarded as an active margin of the Siberian Craton in the Paleasian Ocean or as an Ordovician to Permian island arc based on the subduction-related geochemical signatures of magmatic rocks (Xiao et al., 2010; Chen et al., 2015). Jiang and Nie (2006a, b) reported biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 286 ± 3 Ma and 272 ± 1 Ma and T_{DM2} model ages of 1.45-0.44 Ga for the late Paleozoic granitoids in the Queershan terrane.

Chapter 3. Methodology

One of purposes to perform detailed field observation in the field is to collect ideal samples for structural, paleomagnetic and geochronological studies and to achieve reliable analytic results in the laboratory. Several methods have been applied during this Ph.D. research, such as rock magnetism, anisotropy of magnetic susceptibility, paleomagnetism, U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology, scanning electronic microscope and so on.

3.1 Rock magnetism and magnetic fabric

Before making magnetic analyses, in the laboratory, sampled cores were firstly cut into standard cylindrical specimens, i.e. 2.5 cm in diameter and 2.2 cm in length, for the measurements of Anisotropy of Magnetic Susceptibility (AMS), magnetic remanence and acquisition of Isothermal Remanent Magnetization (IRM). Rock powders or chips were made from representative cores for thermal magnetic and hysteresis curve experiments. Thin sections were prepared for microscopic observations to primarily identify mineral compositions and deformation features of the samples.

In order to identify the carriers of the magnetic remanence and susceptibility, representative samples were chosen for several magnetic mineralogical experiments. Thermomagnetic experiment was carried out using a KLY-3S kappa bridge susceptibility-meter coupled with a CS-3 furnace in the paleomagnetic laboratory of the State Key Laboratory for Mineral Deposits Research (SKLMDR, Nanjing University) and a MFK1 susceptibility meter coupled with a CS-4 furnace at the Paleomagnetism Laboratory of the Institut des Sciences de la Terre d'Orléans (ISTO), Orléans University. The acquisition of IRM was conducted with an ASC IM-10-30 impulse magnetizer and JR-5A and JR-6A spinner magnetometer in the ISTO and SKLMDR. The magnetic hysteretic curves were acquired with Vibrating Sample Magnetometers (VSM) in the paleomagnetic laboratories of the Institute of Earth Sciences, Academia Sinica, and the

Paleomagnetism Laboratory of the Institut de Physique du Globe de Paris (IPGP).

To study the magnetic fabrics, the AMS of specimens were measured using a KLY-3S kappa bridge susceptibility meter in the SKLMDR and a MFK1 susceptibility meter in the ISTO.

3.2 Paleomagnetism

During paleomagnetic sampling, several cautions have been taken to achieve possible field test, including fold test, conglomerate test, contact test, and reverse test.

At least 6 specimens were selected from each site for demagnetization except for some Gubaoquan doleritic dikes from Beishan area since they have small widths (around one meter). Demagnetization was conducted mainly at the Paleomagnetic Laboratory of the SKLMDR (Nanjing) and the ISTO, and also in the Institute of Earth environment, Chinese Academy of Sciences (Xi'an) and the Institute of Earth Sciences, Academia Sinica (Taipei).

Both progressive thermal and alternating field (AF) demagnetization procedures were used to clean the magnetic remanence of specimens. Eleven to fifteen steps with a 50° increment from room temperature to 500°C and a 30° increment from 500°C to 590°C (for magnetite-bearing samples) or 690°C (for hematite-bearing samples) were applied for thermal demagnetization using a magnetically shield oven (ASC, model TD48-SC) at the SKLMDR and a laboratory-built furnace at the ISTO. The AF demagnetization was conducted in 10-13 steps up to 150 mT with intervals varying from 5 to 20 mT using a Molspin demagnetizer hosted in the magnetically shielded room at the SKLMDR and a LDA-3 demagnetizer at the ISTO.

Remanence of specimens was measured with JR-5A and JR-6A spinner magnetometers and a 2G Enterprises Inc. cryogenic rock magnetometer (2G-755) housed in a magnetically shield room. Demagnetization results were plotted on orthogonal vector diagrams (Zijderveld, 1967). Visually identified linear trajectories were used to determine directions of magnetic components by Principal Component

Analysis (PCA), employing a least-square fit comprising three or more demagnetization steps (Kirschvink, 1980), anchoring the fitting lines to the origin where appropriate. Site-mean directions were computed using the PCA-calculated sample directions with Fisher spherical statistics (Fisher, 1953). The paleomagnetic data were analyzed with the software PMGSC (version 4.2 by R. Enkin, unpublished) and software packages offered by Cogné (2003), and Paleomagnetism.org (Koymans et al., 2016).

3.3 Zircon U-Pb geochronology

Zircons were extracted from crushed rock powders using techniques of heavy liquid and magnetic separation. The single zircon grains were handpicked under a binocular microscope and submerged into epoxy resin to form mount disks with a diameter of 1.2 cm. CL imaging was conducted at the State Key Laboratory of Continental Dynamics, Northwest University (Xi'an) with a Quanta 400 FEG scanning electron microscope equipped with a Gatan mini-CL detector (Mono CL3+).

U-Pb isotopic analyses were carried out on an Agilent 7500s inductively coupled plasma mass spectrometry (ICP-MS) coupled with a New Wave 193 μm laser ablation system (LA) installed at the SKLMDR (Nanjing University). Two analyses of an internal zircon standard GJ (608 ± 1.5 Ma; Jackson et al., 2004) and one analysis of an external zircon standard MudTank (732 ± 5 Ma; Black and Gulson, 1978) were made before and after each ten analyses of unknown samples in order to trace instrument stability and to control analytical uncertainty.

Common Pb correction was carried out using the Microsoft Excel[®] embedded program ComPbCorr#3 15G (Andersen, 2002). The program ISOPLOT 3.1 add-in Microsoft Excel[®] was used for age calculation and Concordia plots. Uncertainties are quoted at 1σ for individual analyses and 2σ for weighted mean age calculation.

3.4 $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology

$^{40}\text{Ar}/^{39}\text{Ar}$ dating experiments were conducted at the dating laboratory of the ISTO,

Orléans University on several minerals, including plagioclase, biotite, muscovite, and amphibolite, selected from studied paleomagnetic samples using standard selection criteria for freshness. Standard processes (crushing, hand-picking under a binocular microscope, and ultrasonic cleaning) were applied to achieve the highest possible purity of mineral concentrates. After neutron irradiation for 10 h in the CLICIT Cd-lined slot of Corvallis Nuclear Reactor (Oregon State University, United States) along with the FCS standard (28.02 ± 0.28 Ma, Renne et al., 1998), $^{40}\text{Ar}/^{39}\text{Ar}$ experiments were performed on either single grain (usually biotite, muscovite, and amphibolite) or small-sized concentrates (usually plagioclase) using a high-resolution Helix SFT mass-spectrometer outfitted to a home-built CO₂-laser based extraction system featuring ultra-low argon blanks. Detailed operating conditions can be found in Corti et al. (2019). Ages and isotopes ratios are plotted and tabulated at $\pm 2\sigma$ and were calculated according to Scaillet (2000). Pooled age errors include procedural errors, and decay constants and isotope abundance errors.

3.5 Scanning electronic microscope (SEM) observation

In order to visually characterize main magnetic carrier(s) and unveil whether the paleomagnetic samples have undergone later thermal alteration and/or chemical remagnetization, representative samples were chosen to make polished sections at ISTO, Orléans University for SEM observation. The observation was conducted using a Merlin Compact ZEISS SEM coupled with a Bruker (QUANTAX-XFlash6-30mm²-resolution 129eV) energy dispersive spectroscopy (EDS) system at ISTO, Orléans University. Backscattered SEM images combining the EDS system were used to identify the magnetic minerals. A 15-kV accelerating voltage was set for the instrument during our analysis.

3.6 Paleogeographic reconstruction

The open-source reconstruction software GPlates (www.gplates.org) was used.

Chapter 4. The paleomagnetic study on the Early Permian volcanic and sedimentary rocks and Neoproterozoic dikes in the Yili Block

Although voluminous previous works have been carried out on the tectonic evolution of the Central Asian Orogenic Belt (CAOB; Figure 5-1a), several basic questions are still intensively debated and need further studies. Two of them are presented here: (1) how well constrained is the kinematics of the intracontinental movements (see Section 1.2)? and (2) which are the tectonic affinities and paleogeographic relationships of the microcontinents with Precambrian basements? The overall stratigraphic similarities between the Late Neoproterozoic - Cambrian series exposed on several continental blocks in the CAO (e.g. Ankinovitch, 1962; Figure 4-1b) were variously interpreted. It has been suggested that these continental blocks drifted either from East Gondwana in the Neoproterozoic (Kheraskova et al., 2003), Siberia in the Late Neoproterozoic (Berzin and Dobretsov, 1994), Siberia or Baltica in the Ediacaran (Sengör and Natal'in, 1996), or from the Tarim and/or South China blocks (Levashova et al., 2011).

The late Paleozoic is a key period for studying the amalgamation and the post-orogenic intracontinental tectonics of the CAO (e.g. Wang et al., 2007; Choulet et al., 2011, 2013; Xiao et al., 2015). As the southernmost part of the CAO, the Tianshan Orogen played a major role in the agglomeration and subsequent intracontinental evolution of the CAO. Hence, the relative movements of the three continental blocks in this region (Yili, South Junggar, and Tarim) in late Paleozoic will provide important constraints on the kinematic history of the SW CAO. In addition, a group of Neoproterozoic mafic dikes were found in the north Yili (778~776 Ma; Wang et al., 2014b), which may provide important information about the tectonic affinities and Precambrian paleogeography of the Yili Block. Accordingly, we conducted a paleomagnetic study on the Early Permian volcanic and sedimentary rocks, and the

Neoproterozoic mafic dikes from the Yili Block.

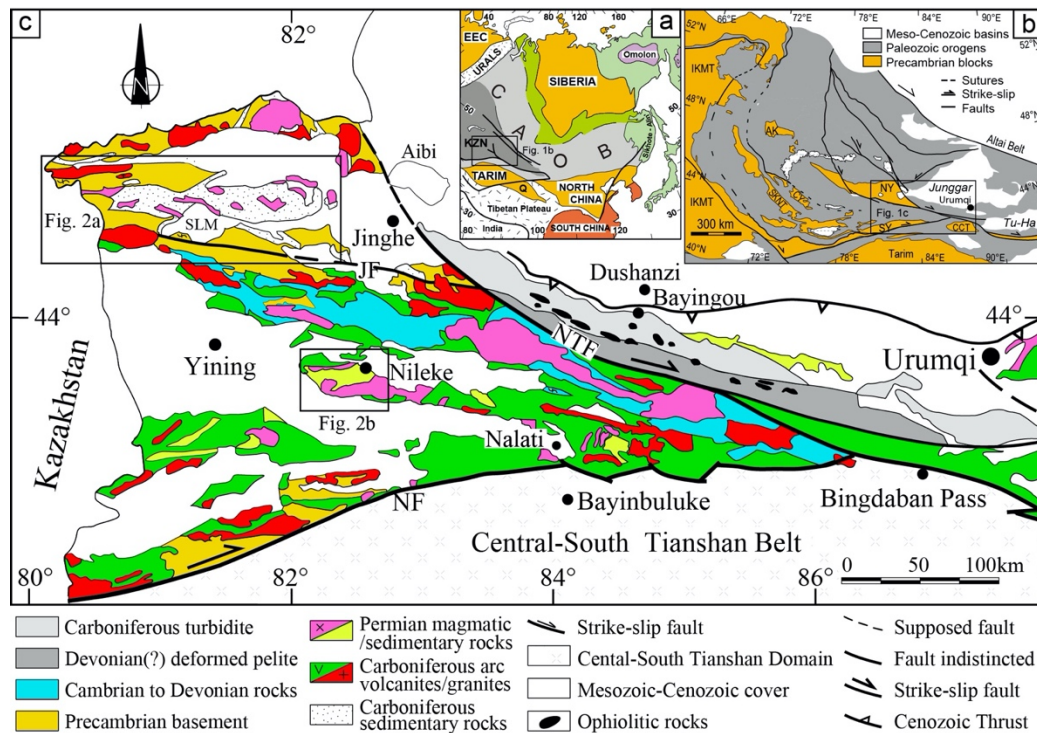


Figure 4-1. (a) Simplified tectonic divisions of the eastern Asia showing the location of the Central Asian Orogenic Belt (CAOB) and its surrounding blocks (after Jahn, 2004; Wang et al., 2014b). EEC = East European Craton; KZN = Kazakhstan; Q = Qaidam. (b) Sketch map of the southwestern part of the CAOB showing the distribution of main microcontinents and accretionary belts. Abbreviations: AK = Aktau, CCT = Chinese Central Tianshan, CY = Chu-Yili, IKMT = Ishim-Kyrgyz Middle Tianshan, NY = North Yili, SKNT = Stepnyak - Kyrgyz North Tianshan, SY = South Yili. (c) Simplified geological map of the Yili Block (modified after XBGMR, 1988a, 1988b). Abbreviations: NTF = North Tianshan Fault, NF = Nalati Fault, JF = Jinghe Fault.

4.1 Geological setting and sampling

4.1.1 Geological setting

In the CAOB, several microcontinents with Precambrian basement are separated by various Paleozoic accretionary complexes (Xiao et al., 2004, 2010; Kröner et al., 2007; Windley et al., 2007). The Yili Block (Figure 4-1b; Wang et al., 2008) is one of these

microcontinents (Gao et al., 1998; Wang et al., 2014a, b) and presents a triangular area in NW China extending westwards into eastern Kazakhstan (Figures 4-1a and b). It is bordered to the north by the North Tianshan Fault in the Chinese North Tianshan, which is a late Paleozoic accretionary complex, and to the south by the Nalati Fault in the Chinese Central Tianshan, which is an early Paleozoic continental arc with a Precambrian basement (Wang et al., 2010, 2018). The North Tianshan Fault and Nalati Fault are two Permian large-scale strike-slip faults that reworked the earlier suture zones (Laurent-Charvet et al., 2002, 2003; Wang et al., 2006, 2009; Charvet et al., 2011).

The Precambrian basement of the Yili Block is mainly exposed along its northern and southern margins; its central part is covered by thick Mesozoic to Cenozoic sedimentary sequences (Figure 4-1c; XBGMR, 1993). The Precambrian basement is mainly composed of Meso- to Neoproterozoic carbonates and amphibolite facies metasedimentary and meta-igneous rocks (XBGMR, 1993; Chen et al., 2000; Hu et al., 2010; Wang et al., 2014a, b; Liu et al., 2014). Early Paleozoic rocks consist of Cambrian to Ordovician carbonates and clastic rocks, and Silurian flysch in the north of the Yining area (Figures 4-1 and 4-2a; XBGMR, 1993; Wang et al., 2014a, b). Late Paleozoic rocks include Devonian terrigenous rocks and arc-type granitoids, and Carboniferous sandstones, limestones intercalated with pyroclastic rocks, volcanic rocks, and coeval granitoids. Permian formations consist of conglomerates and terrestrial sandstones unconformably overlying the older rocks, with widespread Early Permian volcanic and plutonic rocks (XBGMR, 1993; Wang et al., 2009).

In the northern part of the Yili Block, Precambrian rocks exposed around the Sailimu Lake (SW of the Jinghe area) are mainly composed of (1) micaschists, paragneisses, amphibolites, quartzites, migmatites, orthogneisses and gneissic granites belonging to the Wenquan Metamorphic Complex (Figures 4-1 and 4-2a; XBGMR, 1993; Hu et al., 2006, 2010; Wang et al., 2012, 2014a, b; Liu et al., 2014), (2) intensively deformed crystalline limestones and interlayered phyllites of the Jixian System, (3) stromatolite-bearing dolomitic limestones and marbles of the Qingbaikou System, and

(4) non-metamorphic and undeformed sandstones, limestones and interlayered diamictites belonging to the Nanhua and Sinian Systems (XBGMR, 1993; Ding et al., 2009). These Precambrian rocks are unconformably overlain by Cambrian phosphorous sandstones, phosphorites, limestones, mudstones and banded cherts changing upward to Ordovician - Silurian pelitic limestones, cherts, mudstones and black shales (Figure 4-2a; XBGMR, 1988a, 1988b, 1993).

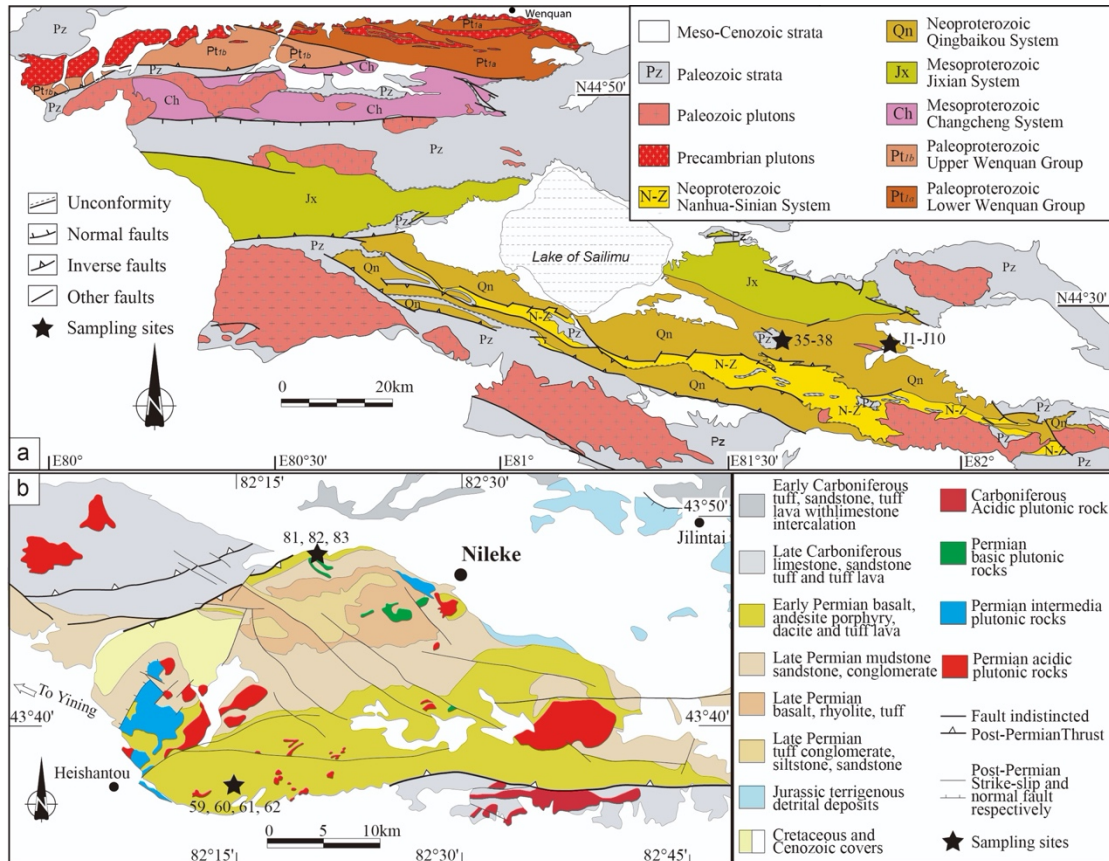


Figure 4-2. Simplified geological maps of the sampling areas: (a) the northern part of the Yili Block (Jinghe area) where sites J1-J10 and 35-38 were sampled; (b) the central part of the Yili Block (Nileke area) where sites 59-62 and 81-83 were sampled.

A group of Neoproterozoic (778~776 Ma) (Wang et al., 2014b) mafic dikes crosscut the carbonates of the Kairtas Group (Qingbaikou System) on the eastern side of the Sailimu Lake (Figure 4-2a; XMBGR, 1988b). The mafic dikes are generally sub-vertical and show an overall ESE - WNW strike with widths varying from ~30 cm to several tens of meters (Figures 4-2a and 4-3a-d). Chilled margins of the dikes and silicification

of the host rocks can be observed to confirm an intrusive contact. A fine-grained texture is often observed in thin dikes and in the marginal facies of the thicker ones, and a coarse-grained texture is commonly present in the inner part of thick dikes. All dikes are composed of variable contents of pyroxene and plagioclase with a subordinate amount of amphibole (Figures 4-3f and h), thus, they belong to gabbros or gabbroic diorites. The intrusion-related thermal aureole is generally more than 50 cm wide (Figure 4-3c). The host rocks consist of argillites, black and white marbles, pelitic limestones and interlayers of carbonaceous shales (Figure 4-3g). These rocks were intensively deformed showing tight folds with sub-vertical bedding and steeply plunging fold axis (Figures 4-3b, d and g). In contrast, no obvious deformation can be recognized in the mafic dikes.

About 10 km to the west of the mafic dikes along the Kusongmuqieke River, a sequence of volcanic and sedimentary rocks crops out in an area about 10 km² (Figure 4-2a). From bottom to top, the sequence is composed of the Lower Permian Wulang Group and the Bakalehe Formation. These Permian strata unconformably overlie the Neoproterozoic Kairtas Group (Figure 4-4a). The Wulang Group consists of calcareous conglomerates, basalts, andesitic basalts, and limestones. The vesicular and amygdaloidal structures can be recognized in basalts and andesitic basalts. The Bakalehe Formation is composed of conglomerates, gray sandstones, and red sandstones unconformably overlying the Wulang Group (Figure 4-4a).

In the Nileke area, central part of the Yili Block, late Paleozoic volcanic and sedimentary rocks are widely exposed (Figure 4-2b). The volcanic and sedimentary sequences can be divided into the Dongtujinhe, Wulang, Xiaoshansayi, Hamisite, Tiemulike and Kashihe Formations (Figure 4-4b). Among them, the Lower Permian Wulang Formation is composed of, from bottom to top, (1) gray tuffaceous sandstones,

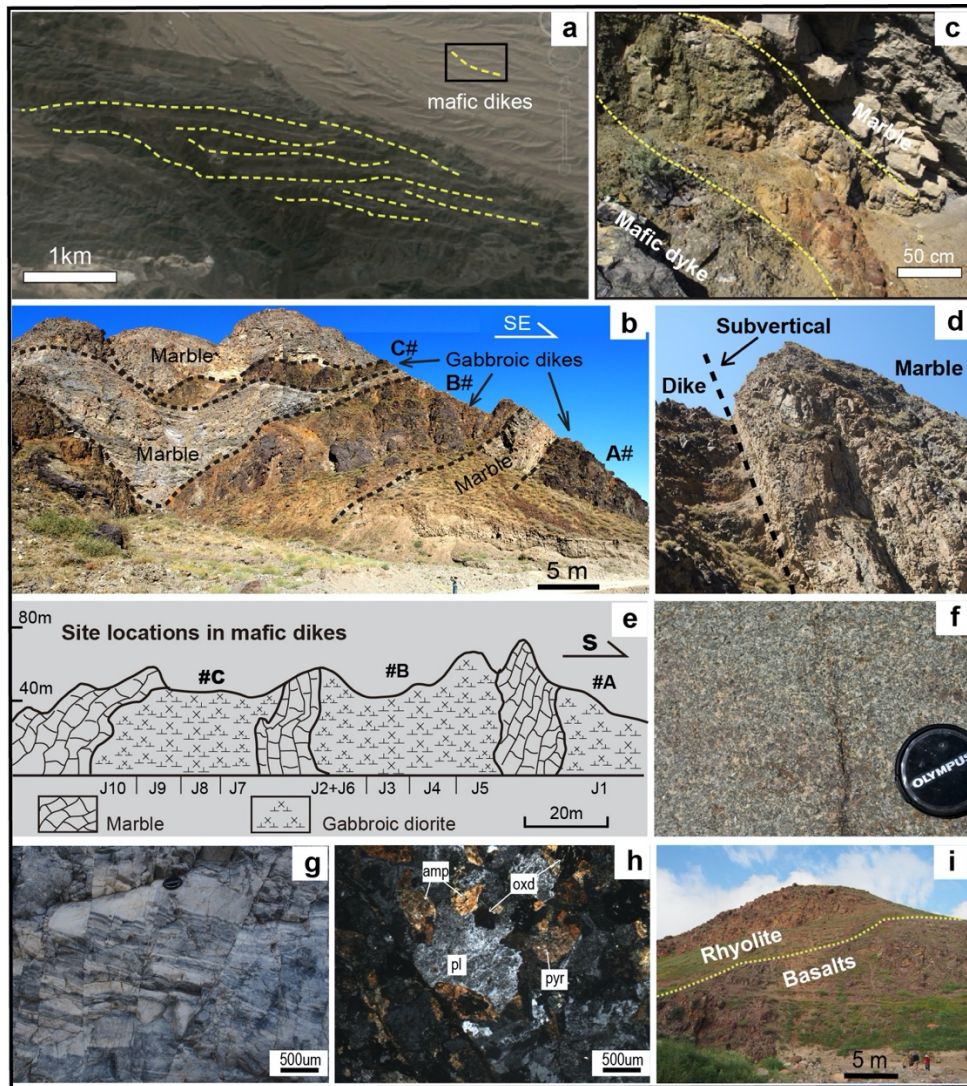


Figure 4-3. (a) Satellite image showing several tens of ESW-WNW-extending mafic dikes intruding the Neoproterozoic meta-carbonates; (b) panorama of three mafic dikes and the host marbles; (c) contact zone between the mafic dike and the host meta-carbonate, a >50 cm-wide interaction zone is characterized by skarnization, cataclastic deformation, and intense secondary alteration; (d) the contact relationship between mafic dikes and host marble suggesting a nearly vertical intrusion; (e) sketch of the Neoproterozoic dikes showing the locations of sampling sites; (f) a coarse-grained texture of gabbroic diorite dike; (g) intensively deformed marbles of the Proterozoic Kairtas Group; (h) photomicrograph of a gabbroic diorite which is composed of subhedral to euhedral amphiboles (amp), pyroxenes (pyr), plagioclases (pl) and minor oxides (oxd); (i) field photograph of basalts sampled from the Nileke area.

andesitic porphyries, and dacite porphyries with tuff breccia, (2) amygdaloidal basalts, sandstones, tuffaceous sandstones, dacite porphyries, amygdaloidal andesitic porphyries, and tuff breccia, and (3) basalts, andesitic porphyries with tuffaceous conglomerates, volcanic breccia, and dacite porphyries (Figure 4-4b).

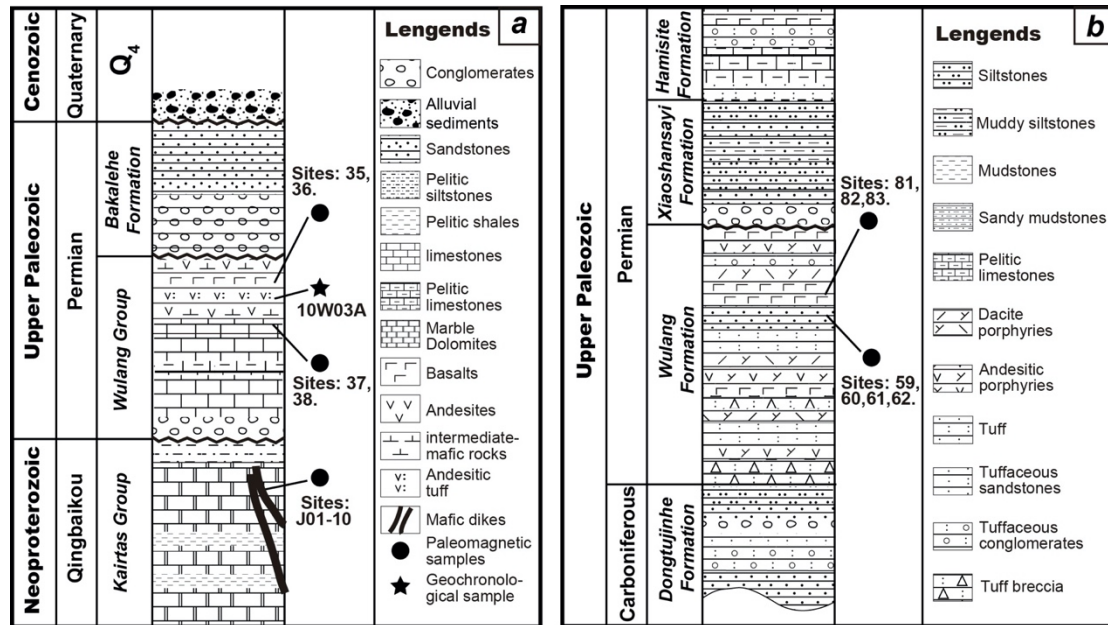


Figure 4-4. Synthetic stratigraphic columns of the sampling areas: (a) the northern part of the Yili Block (the Jinghe area); (b) the central part of the Yili Block (the Nileke area). Solid circles represent the stratigraphic positions of the paleomagnetic samples, and solid star stands for the position of the geochronological sample (10W03A).

4.1.2 Paleomagnetic and geochronological sampling

Tens of mafic dikes can be recognized from satellite images (Figure 4-3a), however, only few of them are accessible due to highly variable topography and very poor road conditions. 84 cores were sampled for paleomagnetic study from 3 thick dikes, namely A# (15m thick, 1 site), B# (42m thick, 5 sites), and C# (30m thick, 4 sites) (Figure 4-3b). These samples were evenly distributed in the dikes, from the chilled margin to the central part (Figure 4-3e). In addition, 40 cores from 4 sites were collected from the Lower Permian volcanic and sedimentary sequences of the Wulang group in the Jinghe area. Two sites were located in the basalts and the other two in the carbonates. An

andesitic tuff sample (10W03A) interlayered in the amygdaloidal basalts and andesitic basalts was also collected for geochronological analysis (Figure 4-4a). Furthermore, 43 cores were sampled from 3 sites of sandstones and 4 sites of basalts in the Lower Permian Wulang Formation of the Nileke area (Figures 4-3i and 4-4b).

More than 6 cores were sampled from each site with a portable gasoline drill, and each core was oriented by magnetic and also by solar compasses whenever possible. The average difference between these two azimuth readings was calculated at $4.2^{\circ} \pm 1.8^{\circ}$, which is used to correct the orientation of samples measured only by magnetic compass. A complete sample list and more details are presented in Table 4-1.

4.2 Magnetic mineralogical investigations

The IRM acquisition curves of samples from mafic dikes show a rapid increase from 0 to 200 mT applied field and the full saturation is reached at about 300mT (Figure 4-5a), indicating the existence of low-coercivity magnetic minerals in the rocks. Thermal magnetic properties of all the measured samples are characterized by irreversible curves of magnetic susceptibility during heating and cooling, suggesting that magnetic mineral transformations occurred during the experiment (Figures 4-5b-e). Figure 4-5b presents the characteristics of the thermal experiments for the dike samples. The progressive decrease in magnetic susceptibility from room temperature to about 400°C may suggest the presence of paramagnetic minerals. Rather significant increase in magnetic susceptibility between 450°C to 570°C probably resulted from mineral transformation from sulfide into magnetite. The rapid decrease of magnetic susceptibility from 570°C to 590°C indicates the existence of magnetite (Dunlop and Özdemir, 1997). No obvious change of magnetic susceptibility occurs after 590°C during the heating. However, a large and rapid increase in magnetic susceptibility is observed during cooling from 590°C to 400°C, which may be the result of the newly formed magnetite. Magnetic susceptibility shows a slightly decrease during the subsequent cooling, indicating possibly a phase change associated with iron sulfides (Roberts et al., 2011). The above

characteristics suggest that magnetite is the main magnetic remanence carrier in the mafic dikes. Magnetic hysteresis loops corroborate this observation (Figure 4-5f).

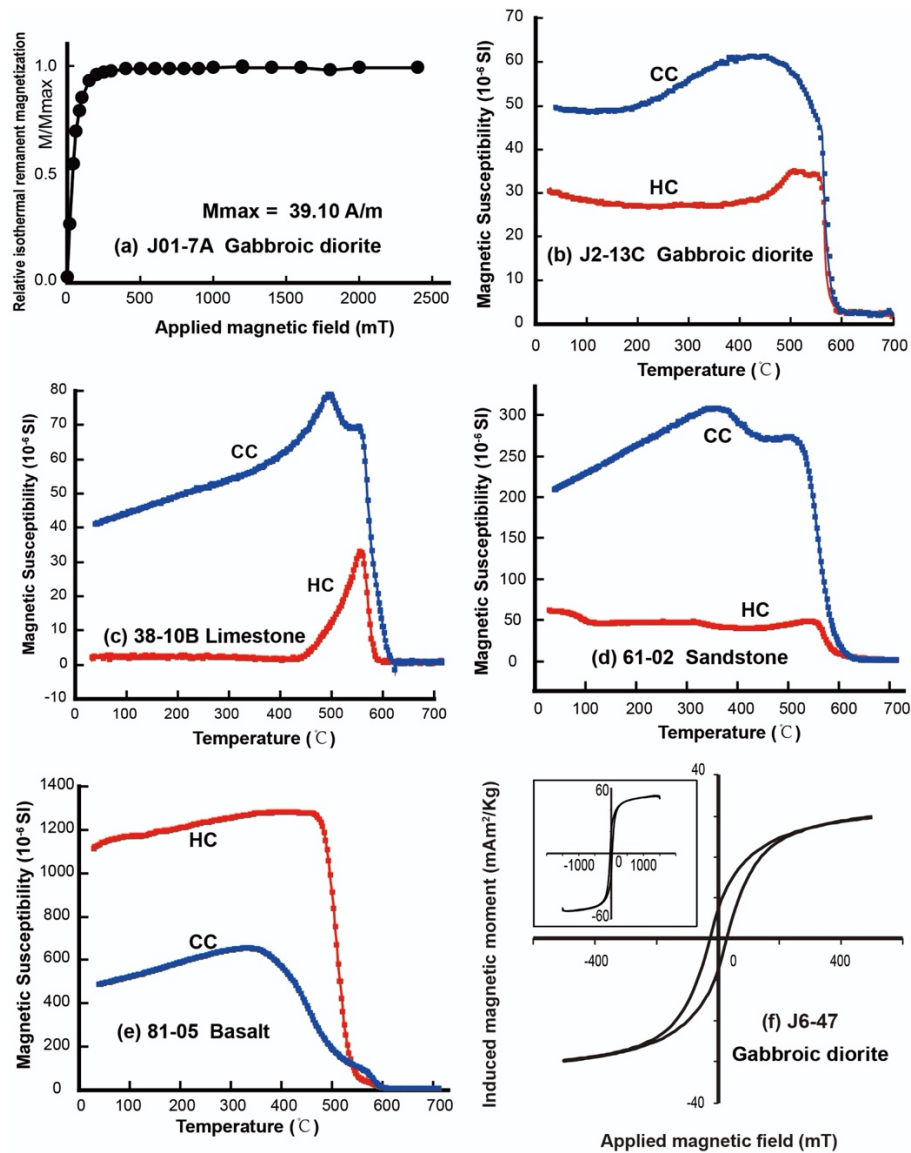


Figure 4-5. Results of different magnetic mineralogical studies indicating that magnetite and minor hematite are the principal magnetic carriers of studied rocks: (a) Isothermal Remanent Magnetization (IRM) acquirement curves, (b-e) Thermo-magnetic (Curie point) experiments, and (f) magnetic hysteretic curve.

Concerning the sedimentary rocks from the Early Permian sequences, a relatively small quantity of magnetite could be probably identified in limestone by a drop of magnetic susceptibility at about 580°C (Figure 4-5c), while the phenomenon of the

newly formed magnetite was clearly observed from 430°C to 580°C (Figure 4-5c). By contrast, magnetite in sandstone is obviously shown in Figure 4-5d with a significant drop at 580°C. In addition, a weak and progressive decrease of the magnetic susceptibility after 580°C can also be recognized and indicates the existence of minor amounts of hematite.

As for the basalts, thermo-magnetic curves prove that magnetite in variable sizes is the principal magnetic remanence carrier (Figure 4-5e). In addition, careful analysis of demagnetization curves (e.g. Figure 4-6h) reveals that variably important magnetic remanence remains after heating at 580°C, indicating the existence of hematite in the basalts.

4.3 Paleomagnetic analysis

For the gabbroic diorite specimens from the mafic dikes, the range of the intensities of the Natural Remanent Magnetization (NRM) is quite large, varying from 1.34×10^{-2} A/m to 2.93 A/m. Most of the demagnetized specimens show a straightforward two-component pattern with a low-temperature component (LTC) removed by ~200°C and a high-temperature component (HTC) isolated from 450°C to 590°C (Figure 4-6a). The results of AF demagnetization usually exhibit a simple single-component pattern (Figure 4-6b). However, there are also some complex demagnetization curves (Figure 4-6c). The LTC directions are dispersed, indicating that they may be originated from viscous remanence. The HTC directions show exclusively reversed magnetic polarity and coherent magnetic directions; therefore, they can be considered as the Characteristic Remanent Magnetizations (ChRMs). A few of them (about 8%) yielded erratic directions by thermal demagnetization due to possibly unstable magnetization or magnetic mineral transformation (Figure 4-6d); therefore, they are excluded from further statistical analysis.

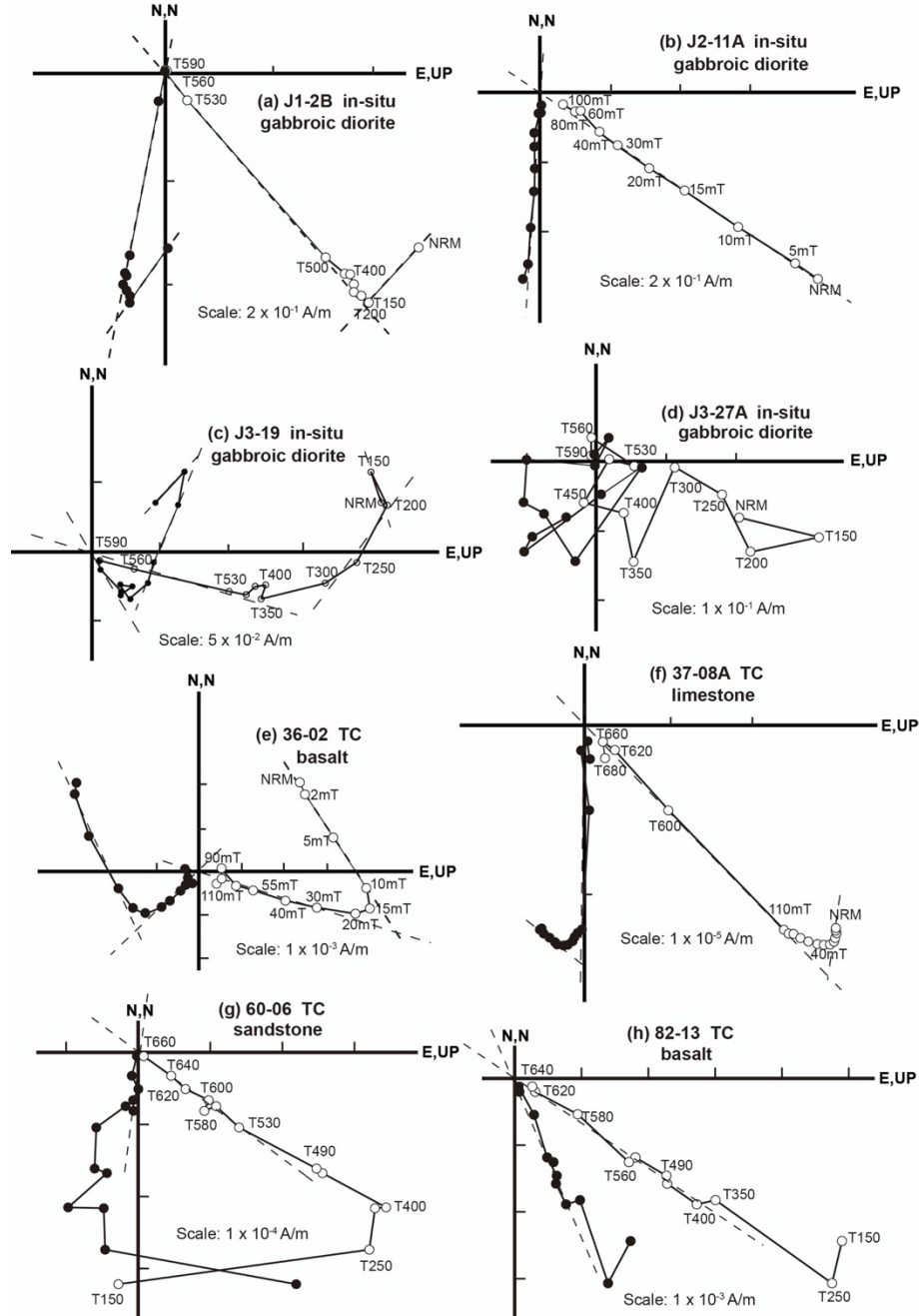


Figure 4-6. Plots of progressive demagnetization for representative samples of Neoproterozoic mafic dikes (a, b, c and d), Early Permian volcanic and sedimentary rocks from the Jinghe area (e and f) and Nileke area (g and h). In-situ and TC (tilting-corrected) stand for geographic coordinates and stratigraphic coordinates, respectively. Solid (open) circles show projections in the horizontal (vertical) plane.

Concerning the volcanic and sedimentary rocks in the Jinghe area, the NRM intensities of the basalts range from 3.63×10^{-3} A/m to 6.15×10^{-2} A/m, and those of

limestone specimens vary from 1.40×10^{-5} A/m to 3.17×10^{-3} A/m. About one-third of specimens yielded erratic directions and are excluded from further directional analysis. In the basaltic specimens, low coercivity components are generally removed by ~ 10 mT and are consistent with the present Earth field; therefore, they are interpreted as the present-day overprints (Figure 4-6e). High coercivity components are generally isolated from 30 mT to 70-100 mT (Figure 4-6e), representing the ChRMs of specimens. As the AF demagnetization cannot totally clean the remanence of the specimens from limestone, a supplementary thermal demagnetization was conducted, and ChRMs are successfully isolated at 660°C (Figure 4-6f).

For the samples from the Nileke area, the NRM intensities of basalts and sandstones range from 6.26×10^{-4} A/m to 1.49×10^{-2} A/m and 1.03×10^{-4} A/m to 1.07×10^{-3} A/m, respectively. Three specimens show erratic directions and are excluded from further analysis. From the remaining specimens, consistent HTC directions are isolated from $\sim 400^\circ\text{C}$ to 560°C , and LTC directions of these specimens are quite dispersed (Figures 4-6g-h).

Mean directions of each site are calculated using Fisher spherical statistics and the results are shown in Table 4-1.

4.4 Zircon geochronology

The analytical data for zircon U-Pb age of the andesitic tuff sample 10W03A are shown in Table 4-2 and plotted in U-Pb Concordia diagram (Figure 4-7b). Thirteen concordant analyses yielded a mean age of 269.4 ± 2.4 Ma (MSWD = 1.3) (Figure 4-7b). The dated zircons are 100 - 150 μm in size and have subhedral to euhedral shapes, and their CL images generally show clear concentric oscillatory zoning that is indicative of an origin from intermediate-acidic magma (Figure 4-7a; Nos. 6, 17, 18, 19, 24, and 30). Ten additional euhedral to subhedral zircon grains (Nos. 1, 3, 5, 7, 8, 9, 11, 12, 13, and 22) have sizes of 50 - 100 μm and are bright with unclear concentric oscillatory zoning (Figure 4-7a; Nos. 3, 9, 12, and 22), and they yielded discordant ages (Table 4-2). The remaining 7 zircon grains also yielded concordant ages, but they should have

derived from different sources. Three zircons yielded relatively young ages (Figure 4-7a; Table 4-2; 258 Ma for zircon No. 15, and 215~230 Ma for No. 25 and 29), and such young magmatic events are unknown in the study area; therefore, these three zircons were likely intermixed into the dated samples from external materials due to contamination. The zircon grain No. 4 yielded an extremely old age of ~2200 Ma with a bright and blurry CL image which is totally different from the others (Figure 4-7a; Table 4-2). Relatively old ages (335~355Ma) were obtained from three subhedral to euhedral zircons (Nos. 26, 27, and 28) with sizes of 100 -150 μm and dark CL images (Figure 4-7a; Table 4-2). These old zircons should be xenocrysts derived from the magma source area or the underlying basement during the magmatism.

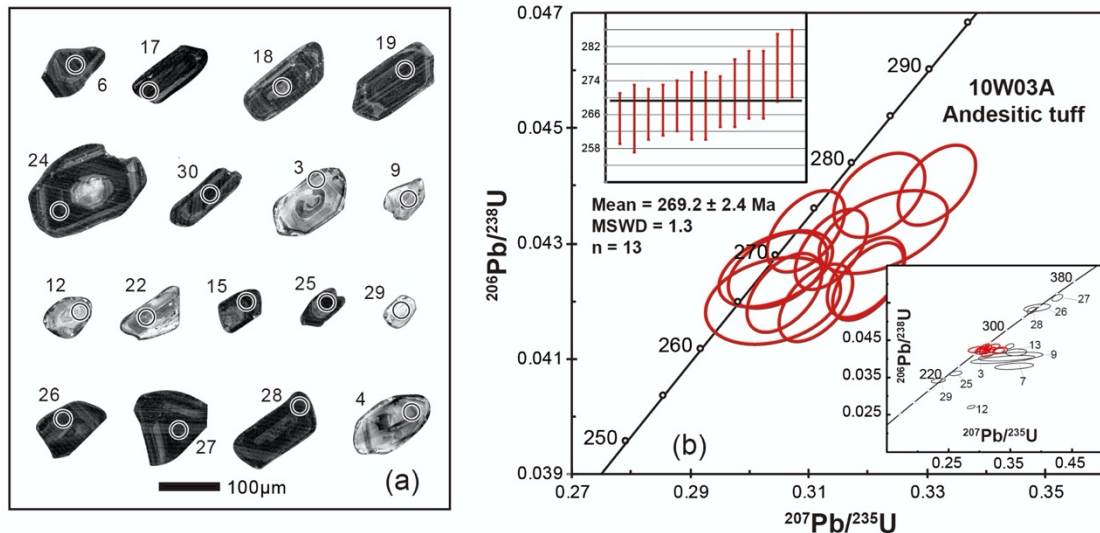


Figure 4-7. (a) Representative cathodoluminescence images and (b) Concordia diagram of dated zircon grains (by LA-ICP-MS U–Pb method) from the andesitic tuff in the northern Yili Block. See Table 4-2 for the isotopic ratios and ages of individual analyses.

4.5 Discussion

Magnetic mineralogical experiments revealed both magnetite and hematite as the principal magnetic carriers in the Neoproterozoic mafic dikes, Early Permian volcanic and sedimentary rocks. Two-component magnetizations have been successfully distinguished through progressive thermal and/or AF demagnetizations. The low-

temperature and low-coercivity components show viscous and unstable magnetic remanences or a present-day overprint, whereas the high-temperature and high-coercivity components decay towards the origin and are considered as the ChRMs. All ChRMs exclusively show a reversed magnetic polarity. They are generally consistent within site and among sites as well (Table 4-1). Before analyzing the tectonic implications of these new paleomagnetic data, their reliability and magnetic remanence age will be discussed considering that the sampling areas experienced multiphase tectonic and magmatic events.

4.5.1 Reliability of the magnetic remanence directions

Concerning the Early Permian (P_1) volcanic and sedimentary sequences, they show exclusive reversed magnetic polarity, which is coherent with the Global Magnetic Polarity scale (Hounslow, 2014); field investigations didn't reveal intensive deformation (only tilting), and no younger magmatism occurred in the study areas after the Early Permian. Therefore, the magnetic remanence directions of the P_1 volcanic and sedimentary rocks can be considered as the characteristic remanence.

The magnetic remanent directions of the Neoproterozoic mafic dikes are considered to be reliable and these dikes were unaffected by deformation (probably only tilting) after their emplacement, or at least after the Early Permian, based on the following arguments: (1) representative thin sections show that all minerals in the mafic dikes are subhedral to euhedral, implying the original magmatic fabrics, and no shape preferred orientation nor dynamic recrystallization could be observed (Figure 4-3h); (2) the subvertical geometry is a characteristic feature for the widespread Neoproterozoic mafic dikes in this region. Numerous dikes of similar ages are found in the NW and NE of Tarim, and these dikes did not experience a post-emplacement folding (e.g. Chen et al., 2004a; Zhan et al., 2007); (3) the orogeny of the Tianshan Belt terminated in latest Carboniferous to earliest Permian (Wang et al., 2008, 2011b; Han et al., 2011), and no significant tectonic event occurred after the Early Permian in the studied area. Post-Permian crustal

shortening in the study area was linked to the Cenozoic India-Asia collision and is characterized by thin-skinned structures (broad and gentle folds) (e.g. Avouac et al., 1993).

4.5.2 Age of magnetic remanence and paleomagnetic pole calculation

Although the andesitic tuff is dated at ~270 Ma, which can be used to constrain the overall formation age of the volcanic and sedimentary sequences of the Jinghe area, and the regional stratigraphic relationship has been characterized for the volcanic and sedimentary sequences in the Jinghe and Nileke areas (XBGMR, 1993), it is necessary to ensure if the magnetic remanence is primary in order to determine its age. Thus, we carried out a careful analysis on the demagnetization curves, and then performed fold tests (Figure 4-8).

By careful analysis on the demagnetization curves, substantial amount of hematite is suggested to exist in the Early Permian sedimentary and basaltic volcanic rocks. For the sedimentary rocks, two components can be separated above and below 580°C, and their difference is generally too small to be significant for a paleomagnetic study (commonly less than 5°, and up to 10° for the extreme case of sample 60-06) (Figure 4-6g). Two possibilities could explain such small difference: (i) experimental errors as the magnetic remanence intensity decreased significantly at high temperature, which would imply that these two components can be considered parallel; (ii) the component carried by hematite is a chemical remanence acquired shortly after deposition, representing the post-depositional remanent magnetization (Butler, 1992; Tauxe, 2010), which is commonly found in many sedimentary environments and is often considered as an accurate recorder of the geomagnetic field at or very close to the time of deposition (Butler, 1992). In the case of our basaltic volcanic rocks showing two indistinguishable remanence components isolated above and below 580°C, two possible processes for remanence acquisition can be proposed: (i) hematite carries a chemical magnetic remanence obtained by oxidation soon after the lava emplacement; (ii) both magnetite

and hematite carry a simple remagnetization. The latter one is inconsistent with the regional geology, i.e., no regional magmatism or high-temperature metamorphism was documented after Early Permian in the study area. In addition, as aforementioned, sedimentary samples in the vicinity or the same sampling sites demonstrate no remagnetization characteristics.

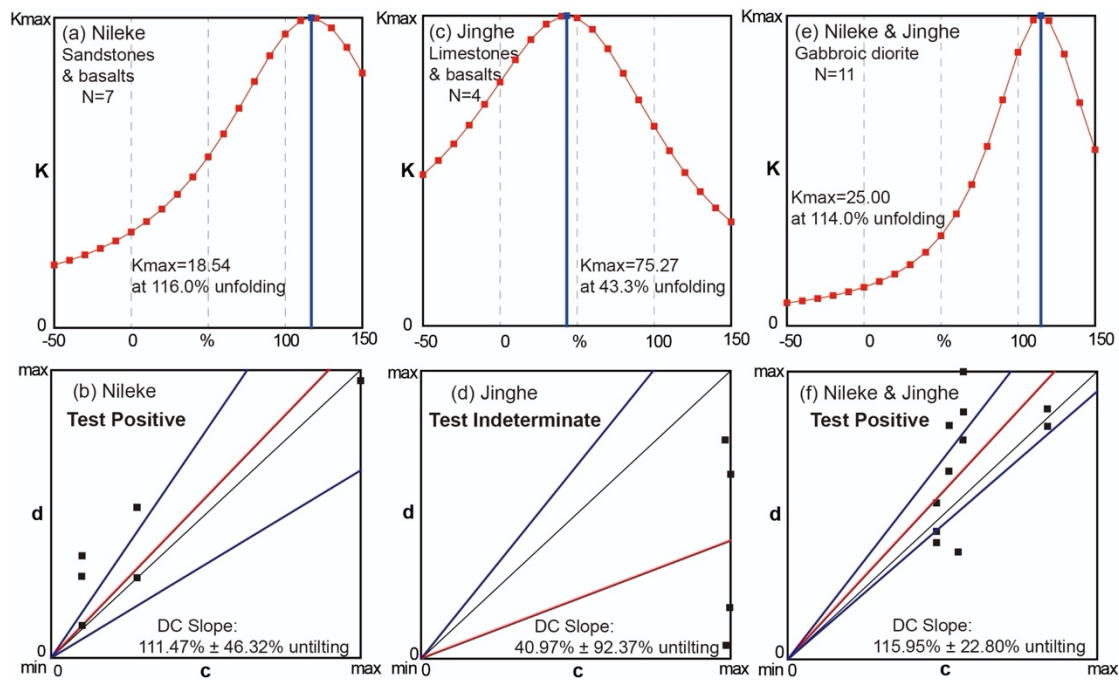


Figure 4-8. Progressive unfolding (PMGSC 4.2 by R. Enkin) and Direction-Correction Test (Enkin, 2003) of the Characteristic Remanent Magnetizations for the Early Permian volcanic and sedimentary samples collected from (a, b) the Nileke area; (c, d) the Jinghe area; (e, f) both Jinghe and Nileke areas.

Stepwise unfolding of the Nileke strata shows that the precision parameter (k) reaches maximum ($k_{\max} = 18.54$) at 116% unfolding with a positive response to the Enkin's fold test (Enkin, 2003; Figures 4-8a-b). The k values are 17.54 after unfolding (k_s) and 5.72 before unfolding (k_g). The variance ratio $f = k_s/k_g = 3.07$ is between 2.69 at the 95% confidence limit and 4.16 at the 99% confidence limit for $N = 7$ (the number of directions of magnetization) (McElhinny, 1964). Thus, a positive fold test at 95% confidence interval is also supported by McElhinny (1964). On the contrary, the k value of the Jinghe strata is highest ($k_{\max} = 75.27$) at 43.3% unfolding (Figures 4-8c-d),

indicating a probable “synfolding” process; however, different variance ratios are calculated at $f_1 = k_s/k_g = 1.22$, $f_2 = k_{\max}/k_s = 1.55$ and $f_3 = k_{\max}/k_g = 1.27$, all these variance ratios are obviously lower than the 95% and 99% confidence limits (4.28 and 8.47, respectively) for $N = 4$ (McElhinny, 1964). Thus, an inconclusive fold test is suggested based on both Enkin (2003) and McElhinny (1964). When the samples from both Nileke and Jinghe areas are analyzed together, the optimal k value ($k_{\max} = 25.00$) is at 114% unfolding with a positive response to the fold test (Enkin, 2003; Figures 4-8e-f), and at 99% confidence interval according to McElhinny (1964), corresponding an $f = k_s/k_g = 22.12/2.99 = 7.40$, higher than the 95% and 99% confidence limits for $N=11$ (2.12 and 2.94, respectively). Statistically, the grouping at 114% is indistinguishable with that at 100%, suggesting a pre-folding remanence.

The fold test on the Jinghe section presents a rather delicate situation, and an original environment on a slope for the volcanic and sedimentary rocks could be one of possibilities, but need additional study to test. The inconclusive fold test is also probably due to the weak bedding variation and small number of sites. In addition, considering that (1) the study area did not experience important deformation nor thermal reactivation events after Early Permian; (2) their remanence magnetization directions are consistent with those of coeval strata from the Nileke area; and (3) there is a good consistency between the pole calculated by tilt-corrected directions and other previous paleomagnetic poles of the Yili Block (Wang et al., 2007), the magnetic remanence of the Jinghe section could be considered as primary. Consequently, the ChRMs of volcanic and sedimentary sequences from both the Nileke and Jinghe areas are more likely primary.

For the Neoproterozoic mafic dikes, we prefer to consider that their magnetic remanence resulted from the Early Permian remagnetization on the basis of the following arguments: (1) a good coherence is observed between the tilt-corrected directions of the P_1 strata and the in-situ directions of the mafic dikes (Table 4-1); (2) in the Jinghe area, about 5-10 km apart from the dikes, unconformable contact can be easily

observed between the P₁ sequence of the Wulang Group and the Neoproterozoic Kairtas Group in which the mafic dikes intruded (XBGMR, 1988a; Figure 4-4a); and (3) intense magmatic activities occurred in the study area during the Carboniferous and Permian. Figure 4-9 shows the age compilations of the Permian magmatic activities in the Yili Block, and the peak of the magmatism is at ~280 Ma. Such massive magmatism and related fluid activities may have easily affected the Neoproterozoic mafic dikes, resulting in their remagnetization.

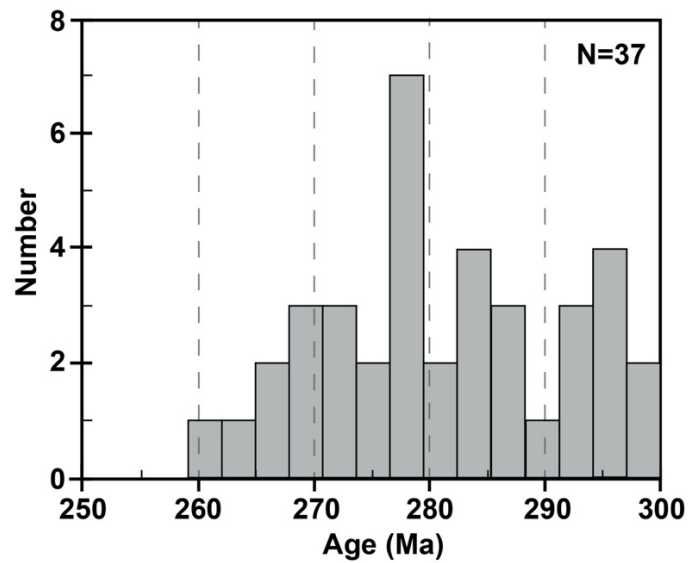


Figure 4-9. The age distribution diagram of the Permian magmatism in the Yili Block (age data from Chen et al., 1994, 2004b, 2007; Duan et al., 2014; Gao et al., 2006, 2009; Gou et al., 2015; Li et al., 2012, 2013, 2015; Liu et al., 2005; Long et al., 2011; Tang et al., 2010; Wang et al., 2004, 2009; Xu et al., 2006, 2013; Yang et al., 2012; Zhang et al., 2008, 2012a, 2012b; Zhao et al., 2003).

However, considering the possibly fast cooling of the studied mafic dikes (Silva et al., 2006), the polarity ambiguity due to the insufficiency of Precambrian paleomagnetic results (Driscoll et al., 2016) and lack of coeval paleomagnetic data from the study block, it is impossible to definitely rule out a primary-ChRMs in the mafic dikes. Thus, further studies are needed to better constrain the age of the magnetic remanence of these Neoproterozoic dikes.

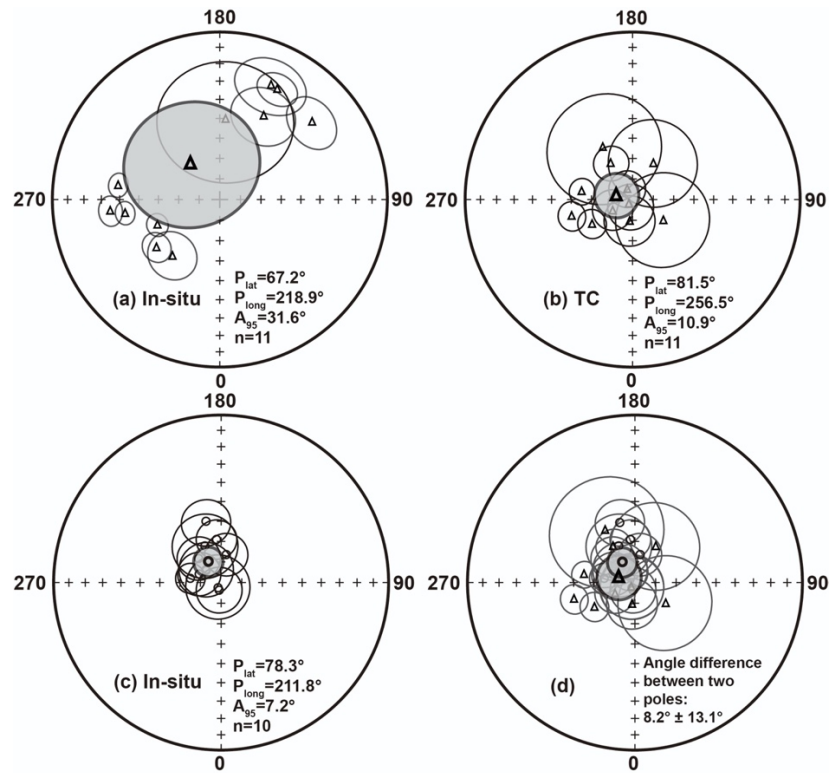


Figure 4-10. Equal-area projections of Visual Geomagnetic Poles (VGPs) from the Early Permian volcanic and sedimentary rocks in (a) geographic and (b) stratigraphic coordinates, and from the Neoproterozoic mafic dikes in the (c) geographic coordinate; (d) comparison of the TC (tilting-corrected) pole calculated from the Early Permian volcanic and sedimentary rocks with the in-situ pole of the Neoproterozoic mafic dikes. Triangle (circle) symbols show poles from volcanic and sedimentary rocks (mafic dikes), and shadowed plots stand for the mean poles. Note: no tilt corrections are made on the VGPs from the Neoproterozoic mafic dikes due to the uncertainty of their magnetic remanence age.

Consequently, we calculated two paleomagnetic poles: one for the Early Permian strata (81.5°N, 256.5°E, $N = 11$ and $A_{95} = 10.9^\circ$) (Table 4-1; Figures 4-10a-b) and one for the Neoproterozoic mafic dikes (78.3°N, 211.8°E, $N = 10$ and $A_{95} = 7.2^\circ$) (Table 4-1; Figure 10c). The poles are statistically indistinguishable with an angular difference of $8.2^\circ \pm 13.1^\circ$ (Figure 10d). Furthermore, since the Early Permian remagnetization of the Neoproterozoic dikes is not yet proved, we will apply only the paleomagnetic pole

calculated from the Early Permian volcanic and sedimentary rocks in the following tectonic interpretations. This is the first Early Permian paleomagnetic pole for the Yili Block.

4.6 Summary

A new paleomagnetic study was conducted on the Early Permian volcanic and sedimentary rocks, and the Neoproterozoic mafic dikes from the Yili Block.

(1) Rock magnetic investigations, including IRM acquisition curves, thermomagnetic experiment, and hysteresis loops, indicated that magnetite is the main magnetic remanence carrier of the Neoproterozoic mafic dikes, magnetite with minor hematite are the magnetic remanence carriers of the Early Permian volcanic and sedimentary rocks.

(2) The andesitic tuff sample collected from the Wulang Group in the Jinghe area has a U-Pb zircon age of 269.4 ± 2.4 Ma, suggesting that the studied volcanic and sedimentary rocks were formed in the Early Permian (dichotomy).

(3) Through careful analyses on the demagnetization curves and fold tests, the Early Permian volcanic and sedimentary rocks probably carry a primary magnetic remanence. Therefore, an Early Permian paleomagnetic pole for the Yili Block is obtained at 81.5°N , 256.5°E , $A_{95} = 10.9^{\circ}$ and $N = 11$ sites.

(4) Although an Early Permian remagnetization is preferred for the Neoproterozoic mafic dikes based on their magnetic remanence directions, field observation, and regional magmatic activities, it is impossible to rule out a primary origin until more data are obtained. Therefore, more future works are needed to understand the paleomagnetic pole acquired from the mafic dikes: 78.3°N , 211.8°E , $A_{95} = 7.2^{\circ}$ and $N = 10$ sites.

Table 4-1. Paleomagnetic sampling and analytical results from the Early Permian volcanic and sedimentary rocks and Neoproterozoic mafic dikes of the Yili Block

Section	Site	Lithology	S-Lat(N°)	S-Long(E°)	Strike (°)	Dip (°)	n/n _o (N)	D _g (°)	I _g (°)	D _s (°)	I _s (°)	k	α ₉₅ (°)	pole		A _{95g} (°)	pole		A _{95s} (°)					
														Lat _g (N°)	Long _g (E°)		Lat _s (N°)	Long _s (E°)						
Volcanic and sedimentary rocks																								
Jinghe	35	Basalt	44°26'57.4"	81°36'27.5"	208	35	6/9	245.1	-59.7	176.7	-61.9	46.2	11.4	43.1	152.4	14.9	87.3	323.6	15.6					
	36						6/9	260.5	-47.9	209.4	-65.9	28.5	14.6	26.3	155.8	15.4	69.5	150.8	21.5					
	37						Limestone	44°26'56.2"	81°37'4.6"	232	48	6/11	262.4	-50.9	198.2	-49.1	72.9	7.9	26.6	152.4	8.8	69.6	210.7	8.5
	38											6/11	275.5	-66.6	172.1	-54.8	68.0	8.2	29.0	129.8	12.2	79.0	297.7	9.8
Nileke	59	Basalt	43°37'13.4"	82°13'53.6"	229	23	5/5	227.4	-49.1	204.9	-40.1	7.7	29.4	50.3	175.9	31.6	60.9	209.4	27.5					
	60	Sandstone	43°38'36.1"	82°14'53.6"	45	40	6/6	163.7	-25.8	186.1	-59.4	98.3	6.8	56.9	292.2	5.4	84.3	207.2	8.9					
	61						6/6	152.8	-23.4	165.6	-61.5	55.1	9.1	50.8	307.2	7.1	79.5	351.9	12.3					
	62						3/6	148.0	-34.4	164.5	-73.2	83.0	13.6	53.6	319.8	11.8	72.0	55.6	23.0					
	81	Basalt	43°48'4.4"	82°21'14.0"	85	49	7/7	180.3	14.7	180.5	-34.3	70.2	7.3	38.7	262.0	5.3	65.0	261.2	6.3					
	82						7/7	168.3	5.4	164.3	-42.4	69.8	7.3	42.3	278.2	5.2	66.9	301.1	7.0					
	83						6/6	168.5	20.6	167.9	-27.6	63.0	8.5	34.6	276.1	6.5	59.1	285.5	6.9					
Mean ₁₁							11	190.2	-39.2			3.0	32.1	67.2	218.9	31.6								
							11			181.1	-53.2	22.1	9.9				81.5	256.5	10.9					
Mafic dikes																								
A#	J1		44°26'23.4"	81°51'45.7"			9/9	179.9	-48.4			58.9	6.8	74.9	262.2	7.2								
B#	J2		44°26'25.9"	81°51'47.5"			10/10	199.2	-62.1			41.5	7.6	76.2	169.5	10.4								
	J3	Gabbroic diorite					6/8	199.5	-51.6			23.8	14.0	70.5	203.9	15.8								
	J4		8/8	186.4	-55.6		26.6	10.9	80.4	229.2	13.2													
	J5		9/9	205.6	-55.6		42.1	8.0	68.9	186.1	9.7													
	J6		8/8	174.7	-62.5		49.4	8.0	86.2	345.4	11.1													

C#	J7	44°26'31.4"	81°51'44.2"	8/8	211.9	-46.6	25.6	11.2	59.7	194.2	11.6
	J8			8/8	183.1	-52.7	38.2	9.1	78.6	248.7	10.5
	J9			8/8	191.7	-50.5	35.0	9.5	74.0	223.0	10.5
	J10			6/8	176.3	-62.0	35.9	11.3	87.1	329.7	15.4
Mean_{no}				10	191.5	-55.3	74.8	5.6	78.3	211.8	7.2

Abbreviations: *S-Lat*(*N*°)/*S-Long*(*E*°), latitude and longitude of sampling sites.

n/no(N), number of specimens in calculation/totally sampled (sites).

D_g, *I_g*, *D_s* and *I_s*, declination (D) and inclination (I) in geographic (g) and stratigraphic (s) coordinates.

k, precision parameter; *α95*, 95% confidence interval.

Lat_g, *Long_g*, *Lat_s*, *Long_s* and *A95*, latitude and longitude of paleomagnetic poles in geographic (g) and stratigraphic (s) coordinates, and 95% confidence interval.

Table 4-2. Zircon LA-ICP-MS U-Pb analytical results of the Early Permian andesitic tuff (sample 10W03A)

Analysis No.	Corrected isotopic ratios						Corrected apparent ages (Ma)						Error correlation	Comments
	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ	²⁰⁷ Pb/ ²³⁵ U	1σ	²⁰⁶ Pb/ ²³⁸ U	1σ	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ	²⁰⁷ Pb/ ²³⁵ U	1σ	²⁰⁶ Pb/ ²³⁸ U	1σ		
10w03a-01	0.05517	0.00158	0.31295	0.00797	0.04114	0.00053	419	65	276	6	260	3	0.43	Discordant
10w03a-02	0.0519	0.00149	0.30449	0.00782	0.04255	0.00054	281	67	270	6	269	3	0.46	
10w03a-03	0.06155	0.00788	0.3379	0.0426	0.03981	0.00088	659	287	296	32	252	5	0.63	Discordant
10w03a-04	0.14109	0.00186	7.97861	0.11723	0.41019	0.00535	2241	23	2229	13	2216	24	0.55	Inheritance
10w03a-05	0.05836	0.00093	0.34882	0.00596	0.04336	0.00057	543	36	304	4	274	4	0.47	Discordant
10w03a-06	0.05198	0.00134	0.30479	0.00683	0.04253	0.00054	284	60	270	5	268	3	0.43	

10w03a-07	0.06801	0.0051	0.35633	0.02588	0.038	0.0007	869	160	309	19	240	4	0.39	<i>Discordant</i>
10w03a-08	0.05726	0.00166	0.33403	0.00957	0.04231	0.00064	502	65	293	7	267	4	0.24	<i>Discordant</i>
10w03a-09	0.06261	0.0078	0.35049	0.04287	0.0406	0.00096	695	279	305	32	257	6	0.47	<i>Discordant</i>
10w03a-10	0.05466	0.00081	0.3316	0.00537	0.04401	0.00057	398	34	291	4	278	4	0.5	
10w03a-11	0.05582	0.00084	0.34448	0.00574	0.04477	0.0006	445	34	301	4	282	4	0.52	<i>Discordant</i>
10w03a-12	0.07676	0.00136	0.28651	0.00536	0.02707	0.00038	1115	36	256	4	172	2	0.44	<i>Discordant</i>
10w03a-13	0.06249	0.00249	0.35962	0.01395	0.04174	0.00074	691	87	312	10	264	5	0.17	<i>Common Pb</i>
10w03a-14	0.05306	0.00074	0.31377	0.00494	0.04289	0.00057	331	32	277	4	271	4	0.55	
10w03a-15	0.05264	0.00169	0.29645	0.00865	0.04085	0.00054	313	75	264	7	258	3	0.47	<i>Contamination</i>
10w03a-16	0.05418	0.00166	0.32266	0.00892	0.0432	0.00058	378	71	284	7	273	4	0.48	
10w03a-17	0.05468	0.00083	0.31966	0.00535	0.04241	0.00057	399	35	282	4	268	4	0.51	
10w03a-18	0.05267	0.00081	0.30599	0.00517	0.04214	0.00056	315	36	271	4	266	3	0.5	
10w03a-19	0.05197	0.00082	0.30955	0.00535	0.04321	0.00058	284	37	274	4	273	4	0.5	
10w03a-20	0.05327	0.00099	0.3219	0.00631	0.04383	0.0006	340	43	283	5	277	4	0.42	
10w03a-21	0.05379	0.00078	0.31125	0.00501	0.04197	0.00055	362	33	275	4	265	3	0.52	
10w03a-22	0.05714	0.00117	0.33198	0.00706	0.04214	0.00059	497	46	291	5	266	4	0.38	<i>Discordant</i>
10w03a-23	0.05075	0.0023	0.29726	0.01267	0.04248	0.00065	230	107	264	10	268	4	0.42	
10w03a-24	0.05481	0.00074	0.31994	0.0049	0.04234	0.00055	404	31	282	4	267	3	0.56	
10w03a-25	0.05255	0.00182	0.26157	0.00829	0.0361	0.0005	309	81	236	7	229	3	0.37	<i>Contamination</i>
10w03a-26	0.05316	0.00254	0.39341	0.01782	0.05367	0.00083	336	111	337	13	337	5	0.45	<i>Inheritance</i>
10w03a-27	0.05483	0.00086	0.42619	0.00733	0.05637	0.00075	405	36	360	5	354	5	0.5	<i>Inheritance</i>
10w03a-28	0.05266	0.00077	0.38481	0.00626	0.053	0.0007	314	34	331	5	333	4	0.52	<i>Inheritance</i>
10w03a-29	0.04955	0.00214	0.23326	0.00951	0.03414	0.00049	174	101	213	8	216	3	0.33	<i>Contamination</i>
10w03a-30	0.05249	0.00179	0.30392	0.00935	0.042	0.00062	307	80	269	7	265	4	0.27	

Chapter 5. The Early Permian Paleomagnetic Pole for the Turpan-Hami Block

Kinematics along the major boundaries between microcontinental blocks and/or terranes is the key to decipher the geodynamic evolution of orogenic belts and to understand mechanisms of transition from subduction-accretion to intracontinental tectonics. Plate boundaries in orogenic belts usually experience long and multi-staged deformation, and geodynamic reconstruction using geologic records often becomes complicated and controversial. Paleomagnetism is a workable method to study these processes, as it can provide quantitative, large-scale constraints on the kinematics of blocks, including paleolatitudinal translation and vertical-axis rotation (Butler, 1992).

Recently, several paleomagnetic studies were conducted in the SW CAOB and provided important constraints for the relative paleogeographic positions of the Yili and South Junggar blocks (Wang et al., 2007; Choulet et al., 2011, 2013; Zhu et al., 2018). However, the paleomagnetic data are still scarce in the Turpan-Hami Block (Figure 5-1b; Cogné et al., 1995; Huang et al., 2004a, b) and remain completely lacking for the Paleozoic. Considering that the Turpan-Hami Block played a pivotal role in the late Paleozoic accretion and the subsequent intracontinental tectonics of the CAOB (Greene et al., 2001, 2005; Wartes et al., 2002; Chen et al., 2015; Wali et al., 2018), Early Permian paleomagnetic data from this area could be of great importance. Therefore, we performed a paleomagnetic study on Early Permian rocks from the Turpan-Hami Basin.

5.1 Geological setting and sampling

The Turpan–Hami basin is bordered by the Bogda Belt and Barkol Tagh to the north, and the Chol Tagh to the south (Figure 5-1b; Greene et al., 2005) and is mainly covered by Meso-Cenozoic sedimentary sequences (Figure 5-1c; XGSC, 2007). The

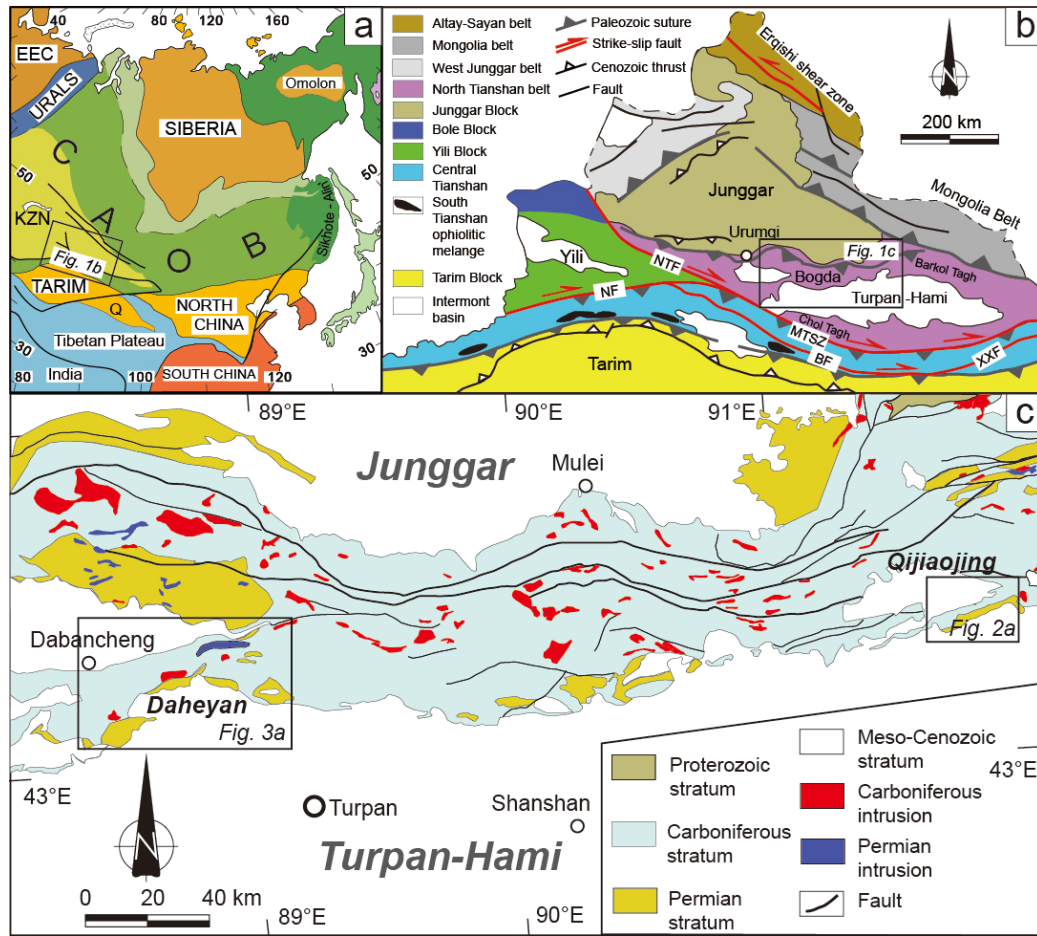


Figure 5-1. (a) Simplified tectonic division of the eastern Asia showing the location of the Central Asian Orogenic Belt (CAOB) and its surrounding blocks (after Jahn, 2004). Abbreviations: EEC, East European Craton; KZN, Kazakhstan; Q, Qaidam. (b) Simplified tectonic division of the Chinese Tianshan area, SW CAOB (after Wang et al., 2007). Abbreviations: BF, Baluntai Fault; MTSZ, Main Tianshan Shear Zone; NF, Nalati Fault; NTF, North Tianshan Fault; XXF, Xingxingxia Fault. (c) Simplified geologic map of the Bogda Belt, northern margin of the Turpan-Hami basin (modified from XGSC, 2007 and Wali et al., 2018).

upper Paleozoic strata are mainly exposed along its northern margin. Our sampling was carried out on Lower Permian strata from both Qijiaoqing and Daheyan areas that are located in the northern margin of the Turpan-Hami basin and are ~240 km apart from each other (Figure 5-1c). Paleomagnetic samples were oriented by both

Late Paleozoic and Mesozoic strata in the Qijiaojing area are composed of, from bottom to top, the Upper Carboniferous Julideneng Formation (C2j), Shaleisaierke Formation (C2s) and Yangbulake Formation (C2y), the Lower Permian Kulai Formation (P1k) and unconformably overlying Upper Jurassic Hongshan Formation (J3h) (Figure 5-2a; XBGMR, 1973). The Kulai Formation is made up by dark gray purple olivine basalts with minor purple quartz porphyries in the upper part, and purple-red siliceous conglomerates, reddish fine sandstones, acidic volcanic breccias, quartz porphyries and andesitic porphyrites in the lower part (XBGMR, 1973).

The lower part of the Kulai Formation was sampled and it consists of basalts, fine- and coarse-grained sandstones, and conglomerates containing rhyolitic clasts. In some geologic maps (XGSC, 2007), these rocks are assigned to be the Lower Permian Aerbasayi Formation (P_{1ae}). The strata are poorly exposed with a total thickness of about one kilometer (Figure 5-2a). They generally strike ENE-WSW (253° - 264°) and dip to north with dip angles varying from 62° to 71°. Due to poor exposure, road conditions and non-continuous outcrop, a relatively small number of outcrops are accessible. Our samples were collected from two sections which are stratigraphically ~200 m apart. From the southern section (QJJ-A), 31 paleomagnetic cores from 3 sites (TH01 to TH03) and a cobble of rhyolite 15T123B were collected (Figures 5-2b and 5-4a); from the northern section (QJJ-B), 32 paleomagnetic cores from 4 sites (TH11 to TH14) and a sandstone dating sample 15T124 were collected (Figures 5-2b and 5-4b). Individual sedimentary beds varying from 3 to 10 m with suitable lithology were selected as sampling sites, and 8 to 13 samples were taken from each site. In total, the strata covered by our sampling is thicker than 130 m (Figure 5-2b). Abundant sedimentary structures, such as load casts and rhythmic layering (Figures 5-4c-d), can be recognized on the outcrops and indicate overturned sequences.

5.1.2 The Daheyan (DHY) area

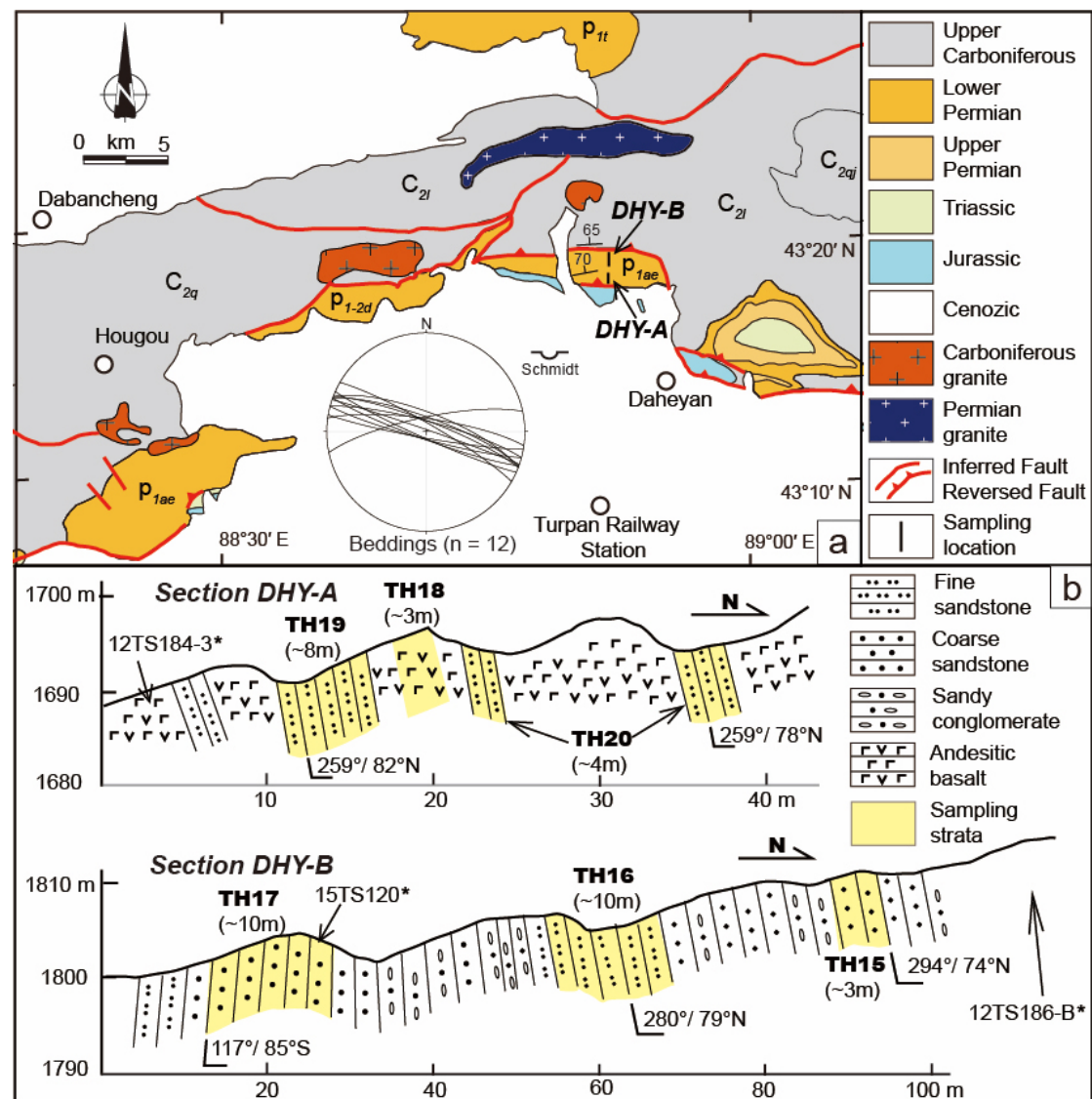


Figure 5-3. (a) Geologic map of the Daheyan area, NW of the Turpan-Hami Basin (modified from XBGMR, 1960, 1988 and XGSC, 2007). DHY-A and DHY-B stand for two sampled sections. (b) Cross-sections of two studied sections in the Daheyan area. Stratigraphic positions of paleomagnetic sampling sites and geochronological samples are shown. Note: geochronological samples labelled with * are after Wali et al. (2018).

The Paleozoic strata from the Daheyan area include, from bottom to top, the Upper Carboniferous Liushugou Formation (C_{2l}), Qijiagou Formation (C_{2qj}) and Qiergusitao

Formation (C2q), the Lower Permian Aerbasayi Formation (P1ae) and Tashikula Formation (P1t), and the Lower-Middle Permian Daheyuan Formation (P1-2d) (XGSC, 2007). These strata are occasionally crosscut by late Paleozoic intrusive rocks (Figure 5-3a). The Aerbasayi Formation is unconformably underlain by or in fault contact with Carboniferous strata, and only exposes in a limited area with a total thickness of around one kilometer (Figure 5-3a); its lower part contains gray purple sandstones, sandy conglomerates and tuffaceous conglomerates, interlayered with quartz porphyries, silicic volcanic breccias and tuffs; its upper part is made up of gray purple andesites, andesitic, rhyolitic porphyries and tuff sandstones (XGSC, 2007).

Two sub-sections were investigated and sampled, and their stratigraphic distance is ~500 m. The strata generally strike sub-E-W (258° - 297°) and dip sub-vertically (74° - 82°) either to the north or to the south (Figures 5-3b and 5-4e). As a result of poor exposure, road conditions and non-continuous outcrop, a relatively small number of sites were collected. In total, 24 paleomagnetic cores from 3 sites (TH18 to TH20) were collected from the southern section (DHY-A) which is composed of basalts and interbedded fine-grained sandstones (Figure 5-3b). Additional 25 paleomagnetic cores from 3 sites (TH15 - TH17) were sampled from the northern section (DHY-B) that consists of gray sandstones and conglomerates (Figure 5-3b). Individual sedimentary beds varying from 3 to 10 m with suitable lithology and a 3 m andesitic basalt flow were chosen as sampling sites, and 7 to 9 samples were obtained from each site. Our sampling covers around 110 m-thick strata (Figure 5-2b). Cross bedding (Figure 5-4f) and rhythmic layering can be observed in clastic rocks, and along with the contact between the basalts and sandstones (Figures 5-4g-h). Overturned sequences are accordingly indicated for most strata except for site TH17. Sub-vertical sandy conglomerates and conglomerates (Figure 5-4i) occur between the studied sections.

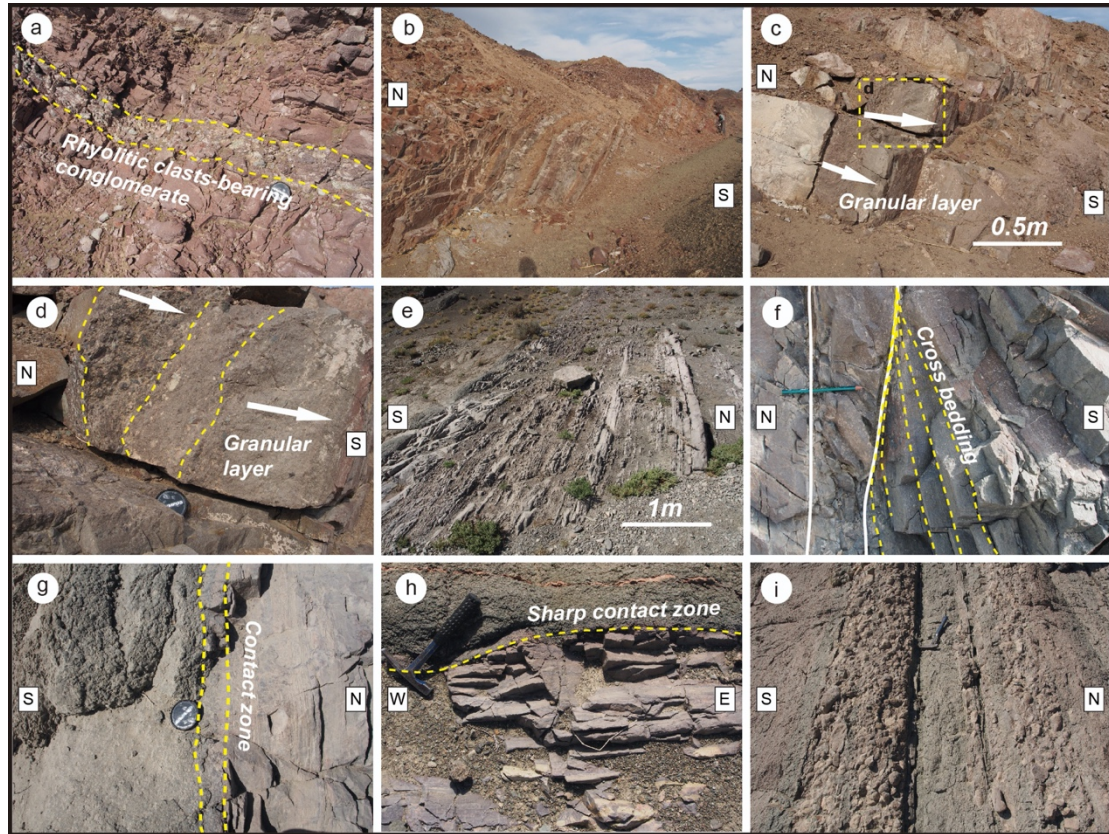


Figure 5-4. Field photographs of Lower Permian strata in the Qijiaojing and Daheyan sections, northern margin of the Turpan-Hami Basin. (a-b) Sandstones and conglomerates including rhyolitic clasts in the Qijiaojing area; (c-d) rhythmic stratification indicating an overturned sequence; (e) sub-vertical beddings of volcanic and sedimentary rocks in the Daheyan area; (f) cross bedding in the sandstone; (g) contact between basalt and underlying baked and slightly hornfelsed sandstone and (h) overlying unbaked sandstone, suggesting an overturned sequence in the Daheyan area; (i) sub-vertical beddings of sandy conglomerates and conglomerates exposed between sections DHY-A and DHY-B.

Wali et al. (2018) provided a zircon U-Pb age for a rhyolite flow (297.9 ± 8.1 Ma) and a maximum deposition age of ~ 297 Ma for a sandstone, which were collected from strata lower than the DHY-B section and from Site H17, respectively (Figure 4b), indicating the oldest formation age at ~ 297 Ma for these strata. In addition, according

to regional stratigraphic data (XBGMR, 1960, 1988; XGSC, 2007), our studied strata belong to the Aerbasayi Formation, and its formation age is reasonably constrained as Early Permian (297 - 273 Ma).

5.2 Rock magnetism and magnetic fabric

In order to recognize the main remanence carriers of the studied samples, rock magnetic mineralogic analyses were conducted, including thermomagnetic experiments, hysteresis measurements, and isothermal remanent magnetization (IRM) acquisition. In addition, magnetic fabric (anisotropy of magnetic susceptibility; AMS) measurements were also carried out to test the reliability of the remanence and to allow possible detection and correction of inclination shallowing bias (Jackson et al., 1991).

5.2.1 Rock magnetism

In light of the characteristics of magnetic mineralogic behaviors, our paleomagnetic samples are divided into three groups (A: TH01 - TH03, B: TH11 - TH14 & TH19 - TH20, and C: TH15 - TH18). For Group A, magnetic susceptibility increases with heating from room temperature to ~400 °C and then decreases with continuous heating up to 680 °C. Rapid decrease happens between 540 °C and 680 °C (Figure 5-5a). The hysteresis loops are wasp-waisted and not fully saturated at a maximum field of 500 mT (Figure 5-5b). These results indicate a mixture of low and high coercivity magnetic minerals (Roberts et al., 1995). The H_c (coercivity of all minerals) and H_{cr} (coercivity of remanence) values are both high, indicating a dominant high-coercivity component. IRM acquisition curves show a rapid increase of IRM below 50 mT and IRM is not saturated at the maximum field of 1.5 T, indicating two magnetic phases contribute to the remanence (Figure 5-5c). Consequently, the magnetic remanence carrier for Group A is suggested to be hematite with minor magnetite.

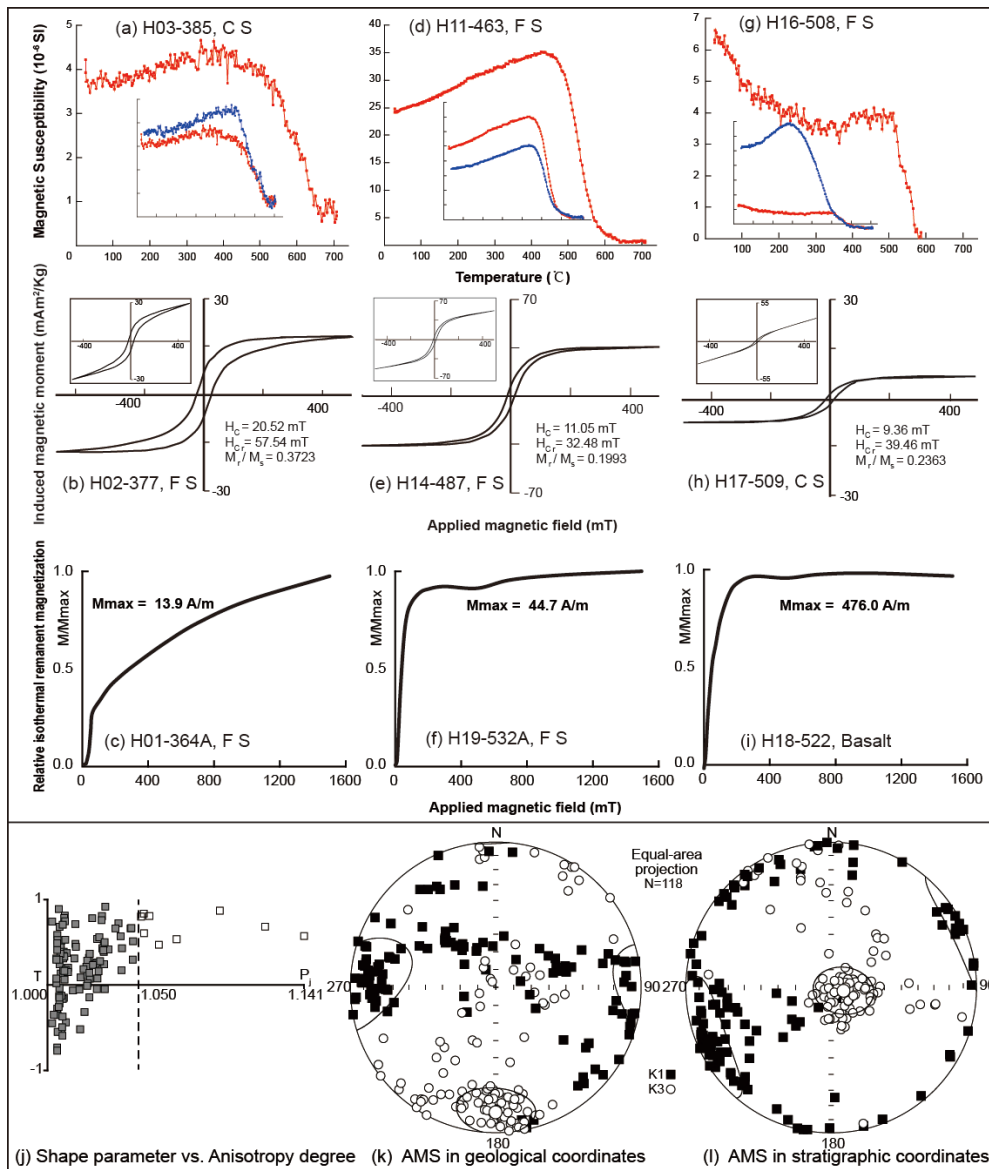


Figure 5-5. Representative rock magnetic analysis results and magnetic fabrics for the samples from the Qijiaojing and Daheyan sections, north of the Turpan-Hami Basin: (a, d and g) results of thermomagnetic experiments. The red and blue curves were acquired during heating and cooling, respectively; (b, e and h) magnetic hysteresis curves; (c, f and i) acquisition curves of isothermal remanent magnetization; (j) plot of anisotropy degree (P_j) versus anisotropy shape (T) of magnetic susceptibility. Open squares represent 9 specimens from Site TH19; (k and l) AMS orientations for 118 specimens in the geographic and stratigraphic coordinates, respectively.

Abbreviations: F S, C S stand for fine-grained sandstones and coarse-grained sandstones.

Samples in Group B show high magnetic susceptibilities, and their thermomagnetic curves are similar to those of Group A (Figure 5-5d). A progressive increase of magnetic susceptibility during heating from room temperature to ~450 °C may be related to mineral transformations of sulfides (Roberts et al., 2011); rapid decrease with heating from 500 °C to 580 °C clearly shows the existence of magnetite (Dunlop and Özdemir, 1997); and further weak decrease when heated up to 640 °C may suggest the presence of minor hematite (Figure 5-5d). The hysteresis loops are narrow, with the saturation below 400 mT (Figure 5-5e), corresponding to low H_c and H_{cr} values, revealing a dominant low-coercivity component. IRM acquisition curves reach 90% saturation at 200 mT, and IRM is not saturated until under a field of 1.5 T (Figure 5-5f). Therefore, magnetite and minor amount of hematite are the principal magnetic carriers for Group B.

Samples of Group C show a progressive decrease of magnetic susceptibility with heating from room temperature to 360 °C (Figure 5-5g), which agrees with the typical magnetic behavior of paramagnetic minerals (Curie Law; Tauxe et al., 2010); and subsequent increase with heating up to 520 °C could be due to mineral transformations of sulfides; the sharp decrease with continuous heating thereafter from 520 °C to 590 °C suggests the presence of magnetite (Figure 5-5g). The hysteresis loops are saturated below 200 mT; their low H_c and moderately high H_{cr} values imply that the magnetite grains are small (Figure 5-5h). IRM acquisition curves saturate at ~200 mT (Figure 5-5i), which consistently proves that magnetite is the major magnetic carrier for the samples of Group C.

5.2.2 Magnetic fabric (AMS)

The magnetic fabrics were measured on a total of 118 specimens from both Qijiaojing and Daheyan areas. The magnetic susceptibilities of four basalt specimens vary from 2.64×10^{-2} to 5.02×10^{-2} SI and those of other specimens range from 2.45×10^{-4} to 9.38×10^{-3} SI. The AMS results of all specimens show a magnetic fabric

characterized by oblate ellipsoids (Figure 5-5j). Most specimens show PJ (anisotropy degree of AMS) values below 1.05, ranging from 1.002 to 1.047, with a mean of 1.012 (Figure 5-5j); and only nine specimens from Site TH19 reveal PJ values varying from 1.051 to 1.141 with a mean of 1.072.

Overall speaking, the in-situ Kmax axis (the maximum axes of the ellipsoid used to represent the magnetic anisotropy; magnetic lineation) shows a near E-W orientation ($272.9^{\circ} \pm 31.9^{\circ}$; Figure 5-5k). After tilt correction, Kmin axes (the minimum axes of the magnetic anisotropy ellipsoid; pole of magnetic foliation) cluster to be perpendicular to horizontal plane and Kmax axes scatter horizontally (Figure 5-5l), indicating that the oblate magnetic fabric is mainly related to the depositional process.

5.3 Paleomagnetism

At least seven specimens were selected from seven samples from each site for demagnetization. Both thermal and alternating field (AF) demagnetizations were carried out to isolate characteristic remanent magnetizations (ChRMs). Tilt corrections for overturned sequences were achieved by restoring the sequences to their original horizontal attitude (more specific, use opposite measured strike direction and supplementary angle of measured dip angle to conduct corrections in a paleomagnetic software).

ChRMs are successfully isolated from ~85% of measured specimens; the remaining ~15% of the specimens show erratic demagnetization results due to weak magnetic remanence or thermal transition during heating, therefore, they were excluded from further analysis. Thermal and AF demagnetization diagrams of twin specimens from the same core are comparable (Figures 5-6a-b). Nevertheless, thermal demagnetization is generally more effective than AF one as the latter cannot completely isolate the remanence of hematite-bearing specimens (Figure 5-6a).

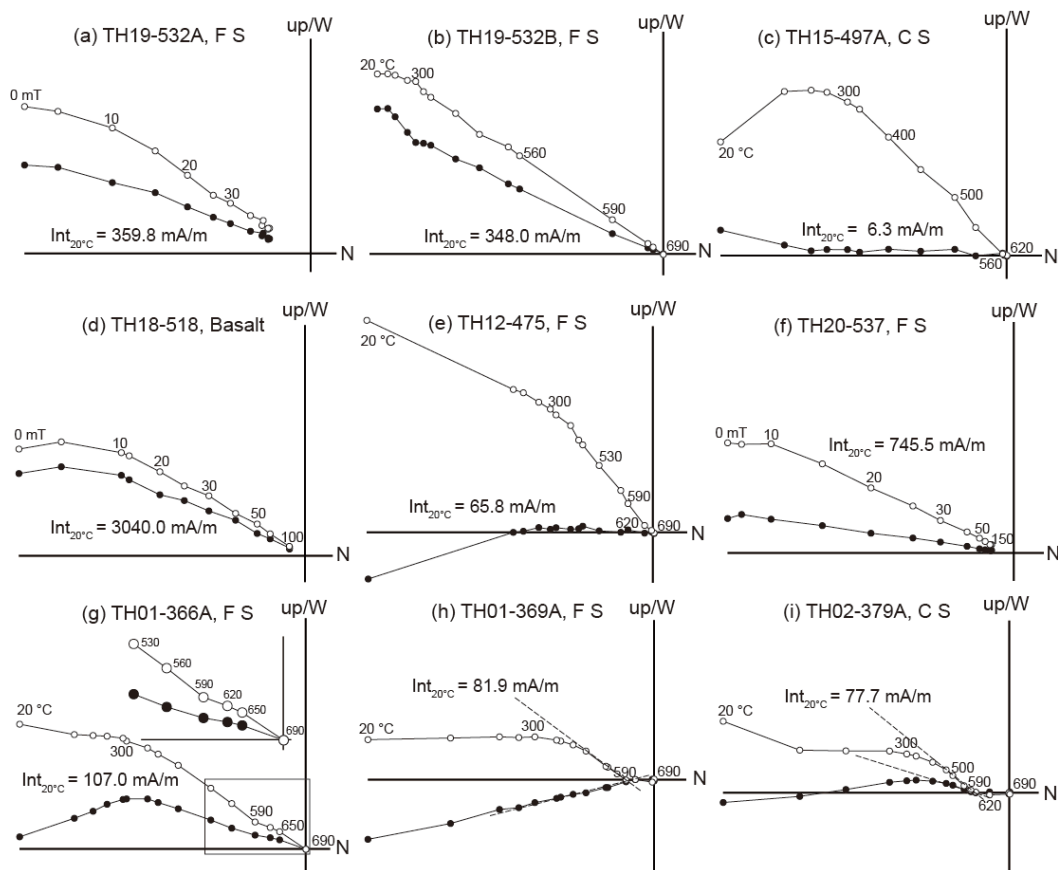


Figure 5-6. Representative demagnetization diagrams for sedimentary and volcanic rocks collected from the Qijiaojing and Daheyan sections, north of the Turpan-Hami Basin. All demagnetization plots are displayed after bedding tilt corrections. Solid (open) symbols stand for the projection of vector end-points on the horizontal (vertical) plane. Abbreviations: F S, C S stand for fine-grained sandstones and coarse-grained sandstones.

Depending on whether or not hematite is the main remanence carrier of specimens, the demagnetization results can be classified into two groups. For the specimens in which magnetite is the main remanence carrier, two components are easily isolated: the low-temperature or low-coercivity components are commonly removed at 200-250 °C or 10-20 mT; and the high temperature or high-coercivity components (ChRMs) are isolated from at least four successive steps in the interval of 400-590 °C or 20-100 mT (Figures 5-6c-d).

For the specimens with hematite as a main remanence carrier, the low-temperature or low-coercivity components are also unblocked at 200-250 °C or 10-20 mT. However, their ChRMs are complicated and three cases are recognized. (1) Most specimens show ChRMs isolated from at least five steps in the interval 400-690 °C or 20-150 mT (Figures 5-6e-f). (2) A few specimens reveal a slight direction difference between the components isolated in the 400-590 °C and 590-690 °C intervals (Figure 5-6g). Their demagnetization curves from 400 to 590 °C are straight, indicative of a single component; and those from 590 to 690 °C are curvilinear, likely due to a superposition effect of two components, similar to descriptions in Butler (1992). Because their remanence directions are almost the same, the ChRMs were analyzed by using 450-690 °C interval with forcing the fitting line towards the origin. (3) The remaining specimens demonstrate two significantly different high-temperature components isolated in 400-590 °C and 590-690 °C intervals (Figures 5-6h-i). The remanence directions of the 400-590 °C component are consistent with other ChRMs and can be considered as the ChRMs of these samples. Whereas, the 590-690 °C components generally show a shallow inclination and a declination of $\sim 180^\circ$, which are completely different from the ChRMs acquired from other specimens, thus, they are regarded as a possible secondary remanence and their importance will be discussed later.

All the isolated ChRMs yield a reverse polarity. Estimated site-mean directions are listed in Table 5-1. In the Qijiaojing area, the ChRMs of the two sections are consistent, resulting in a tilt-corrected mean remanence direction of $D \pm \Delta D = 179.6^\circ \pm 3.9^\circ$, $I \pm \Delta I = -39.7^\circ \pm 5.0^\circ$ ($n = 63$, $k = 26.0$, $\alpha_{95} = 3.6^\circ$) (Figure 5-7a; Table 5-1). Two sections in the Daheyan area show significantly different mean remanence directions (Figure 5-7c). The mean direction of the section DHY-A is calculated as $D \pm \Delta D = 205.3^\circ \pm 2.7^\circ$, $I \pm \Delta I = -33.0^\circ \pm 4.1^\circ$ ($n = 25$, $k = 113.3$, $\alpha_{95} = 2.7^\circ$; Figure 5-7c; Table 5-1), whereas the one obtained at the section DHY-B is $D \pm \Delta D = 178.3^\circ \pm 4.4^\circ$, $I \pm \Delta I = -50.6^\circ \pm 4.2^\circ$ ($n = 35$, $k = 48.4$, $\alpha_{95} = 3.5^\circ$; Figure 5-7c; Table 5-1). The isolated low-temperature or low-coercivity components are more concentrated in geographic coordinates (mean

direction: $D = 6.5^\circ$, $I = 63.5^\circ$, $\alpha_{95} = 1.8^\circ$, $n = 80$), which is indistinguishable with the present Earth field ($D = 2.0^\circ$, $I = 63.8^\circ$; PaleoMac; Thébault et al., 2015), than in stratigraphic coordinates (mean direction: $D = 172.2^\circ$, $I = 6.3^\circ$, $\alpha_{95} = 5.3^\circ$, $n = 80$) (Figure 5-8).

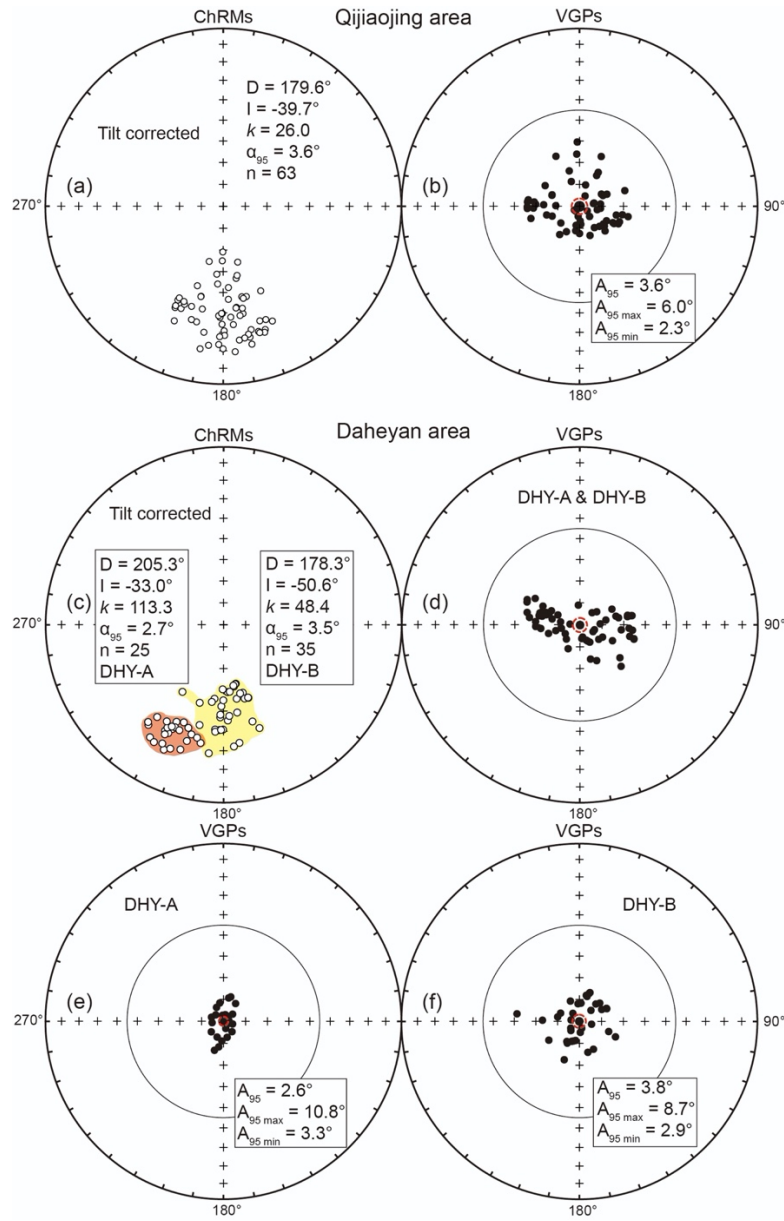


Figure 5-7. Equal-area projections of the isolated ChRM directions in stratigraphic coordinates from the Qijiaoqing section (a) and Daheyuan (DHY-A & DHY-B) section (c), and the corresponding VGPs for the Qijiaoqing section (b), Daheyuan (DHY-A & DHY-B) section (d), DHY-A section (e) and DHY-B section (f), respectively. Cut-off criterion of 45° is used after Deenen et al. (2011).

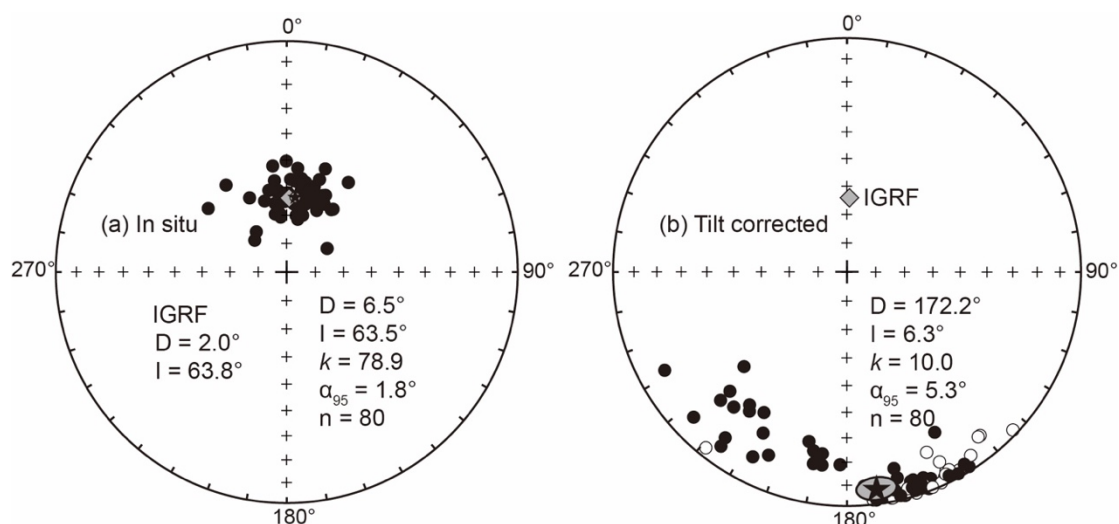


Figure 5-8. Equal-area projections of the low-temperature and low-coercivity components in (a) geographic and (b) stratigraphic coordinates, which were isolated from the Early Permian strata in the Qijiaojing and Daheyan areas, northern margin of the Turpan-Hami Basin. Note: gray quadrangle stands for the present magnetic direction in the sampling areas according to International Geomagnetic Reference Field (IGRF, 2013).

5.4 Zircon U-Pb Geochronology

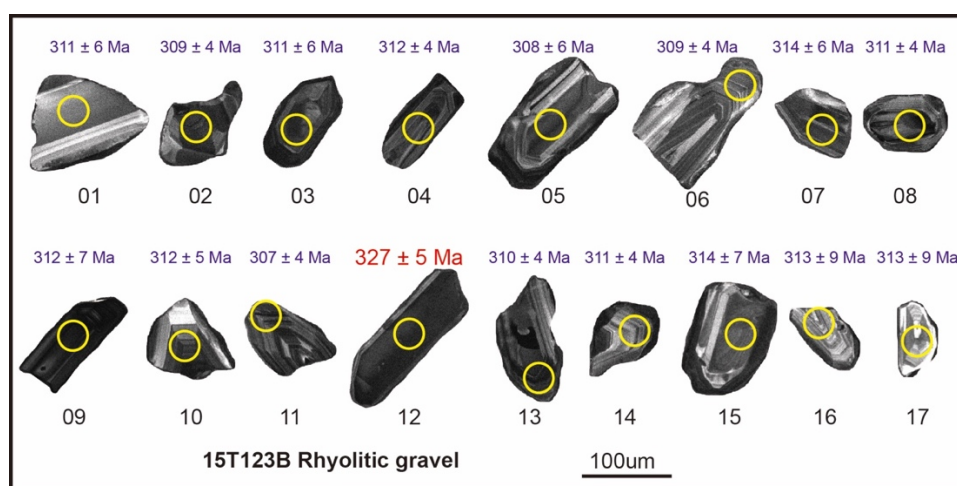


Figure 5-9. Cathodoluminescence images and corresponding $^{206}\text{Pb}/^{238}\text{U}$ apparent ages of the analyzed zircons selected from the rhyolitic gravel sample 15T123B. The older apparent age of zircon No.12 is shown in red.

Two samples for zircon U-Pb dating were prepared and analyzed at the LA-ICP-MS Laboratory of the State Key Laboratory for Mineral Deposits Research (SKLMDR, Nanjing University). The cathodoluminescence (CL) images of dated zircons and analytical results are reported in Figures 5-9, 5-10 and Table 5-2.



Figure 5-10. Cathodoluminescence images and corresponding $^{206}\text{Pb}/^{238}\text{U}$ apparent ages for analyzed detrital zircons selected from the sandstone sample 15T124. The youngest $^{206}\text{Pb}/^{238}\text{U}$ apparent ages are shown in red.

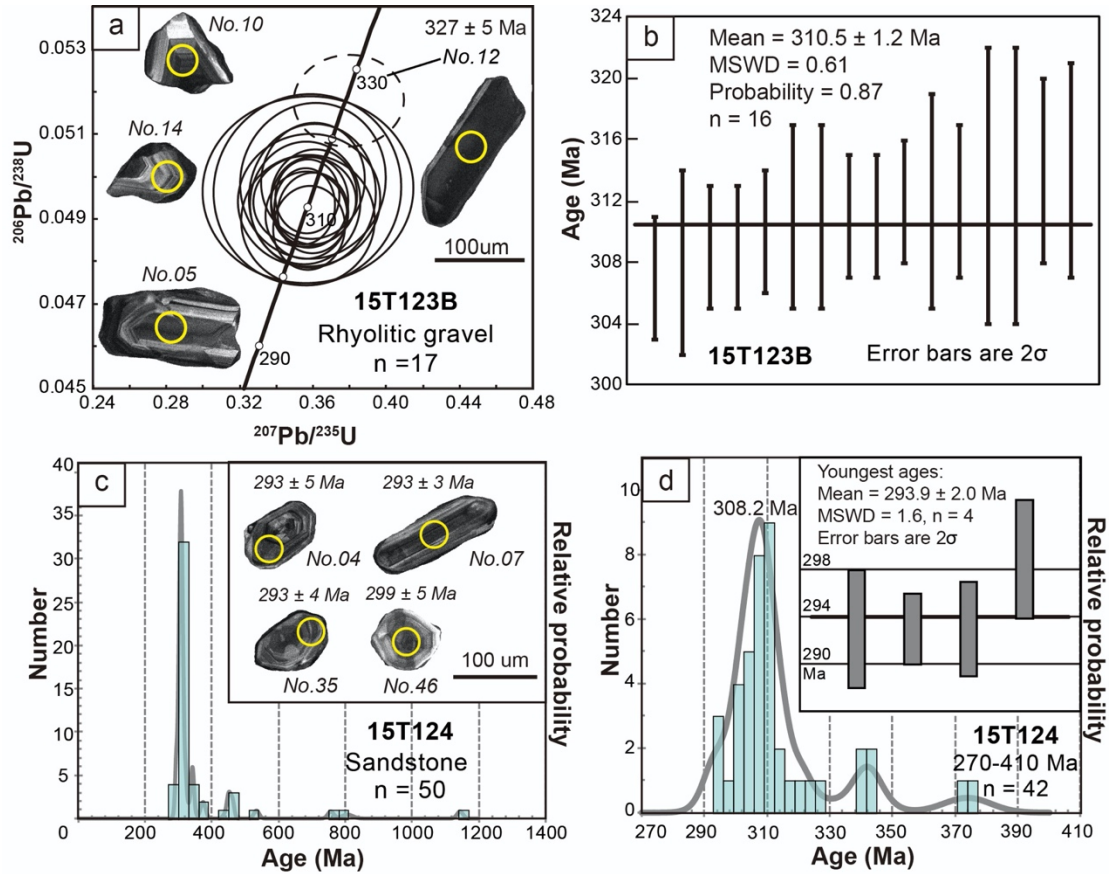


Figure 5-11. (a) Concordia diagram of zircon U-Pb dating and cathodoluminescence images of representative zircons for the rhyolitic cobble sample 15T123B from the conglomerate of the Qijiaoing area; (b) weighted mean zircon U-Pb age for sample 15T123B calculated from 16 out of 17 individual ages; (c) histogram and relative probability curve for detrital zircon U-Pb ages of the sandstone sample 15T124 from the Qijiaoing area, the cathodoluminescence images of the youngest zircons are shown; (d) probability density function plot of sample 15T124 shown in 270-410 Ma window, and the weighted mean age calculated from the youngest ages.

For the rhyolitic cobble sample 15T123B collected from the Kulai Formation, the dated zircons are 80-200 μm in size and have subhedral to euhedral shapes with length/width ratios of 1.5-3, and their CL images generally show clear oscillatory zonings which indicate an origin from a siliceous melt (Figures 5-10 and 5-11a). Seventeen concordant zircons yielded sixteen consistent apparent ages, and a weighted

mean age is calculated at 310.5 ± 1.2 Ma with MSWD = 0.61 (Figures 5-11a-b). One exceptional older age (327 ± 5 Ma) was obtained from zircon No.12 (Figure 5-11a; Table 5-2), whose CL image is plain dark without oscillatory zoning, and this zircon should have derived from different source. Therefore, the depositional age of this formation should be younger than about 310 Ma.

Detrital zircons from the sandstone sample 15T124 are 50-200 μm in size and have sub-rounded to sub-euhedral shapes with length/width ratios of 1-3 (Figure 5-10). Their CL images are generally black or bright with concentric oscillatory zonings; however, some zircons have plain dark CL images and no oscillatory zoning is visible (Figure 5-10). Fifty concordant ages ranging from ~ 293 Ma (Nos. 4, 7, 35 and 46; Figure 8c) to 1145 Ma (No. 50) are acquired, which define a uniform peak at ~ 308 Ma (Figures 5-11c-d; Table 5-2). The weighted mean age of the youngest four zircons is calculated at 293.9 ± 2.0 Ma (Figure 5-11d). Thus, the maximum depositional age of the sandstone could be approximated as 294 Ma.

5.5 Discussion

5.5.1 Inclination shallowing test

Inclination shallowing can occur in sedimentary rocks due to compaction and/or rock-magnetic effects and has been frequently documented in previous studies (e.g., Kodama, 2012). A test of possible inclination shallowing is therefore indispensable for paleomagnetic studies on sedimentary rocks in order to obtain reliable tectonic implications (e.g., Huang et al., 2015). Two most widely used correction methods are the elongation/inclination (E/I) correction of the ChRM distribution (Tauxe and Kent, 2004; Tauxe et al., 2008) and the one based on anisotropy of magnetic susceptibility or remanence (Jackson et al., 1991; Kodama, 2009).

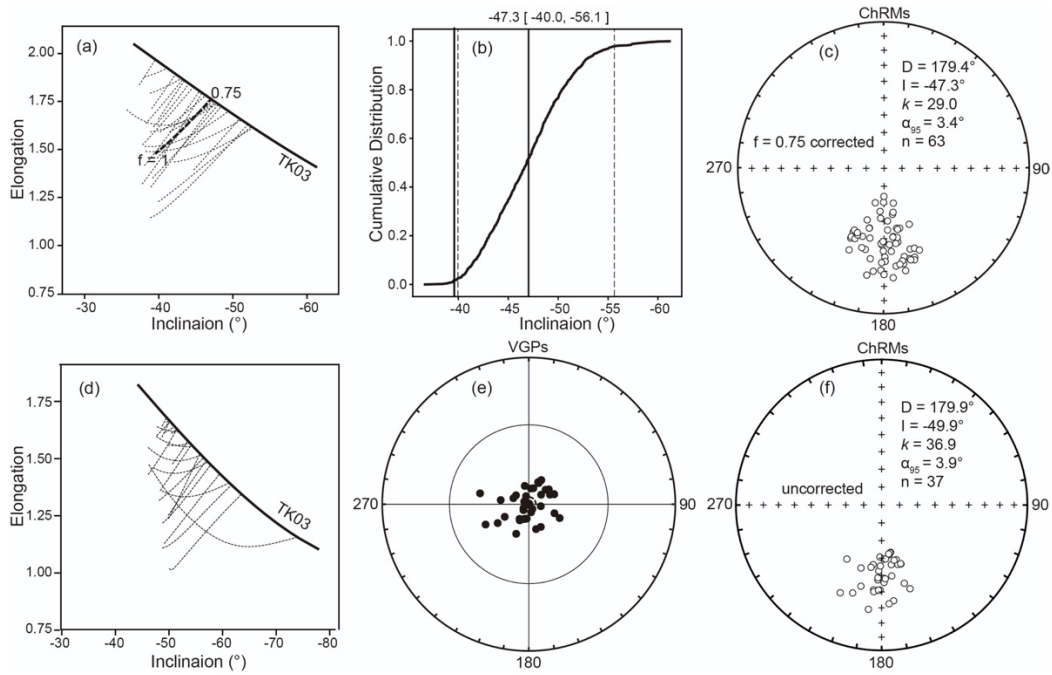


Figure 5-12. E/I corrections on the paleomagnetic specimens from the Qijiaojiang area (a-c) and Daheyan area (d-f). (a) Plot of elongation versus inclination for the TK03.GAD model (solid line) and the data from the Qijiaojiang area (dashed line). Twenty bootstrapped datasets are also shown. The crossing points represent the inclination/elongation pair most consistent with the TK03.GAD model. (b) Cumulative distribution of crossing points from 1000 bootstrapped datasets. The dashed lines indicate the 95% confidence interval. (c) Equal-area projections of corrected ChRM directions in stratigraphic coordinates. (d) No intersection was found between the TK03.GAD model and the data from the Daheyan area. (e, f) Equal-area projections of VGPs and ChRMs, respectively, for the data from the Daheyan area in stratigraphic coordinates.

The E/I correction is based on the paleosecular variation models TK03 (Tauxe and Kent, 2004). Since the Daheyan and Qijiaojiang areas are ~240 km apart, their E/I corrections were separately conducted. For the strata in the Qijiaojiang area, according to the Deenen et al. (2011) criteria, the distribution of the ChRM directions reflects the contribution from paleosecular variation (PSV) of the geomagnetic field (i.e. A95 min

$= 2.3^\circ < A95 = 3.6^\circ < A95 \text{ max} = 6.0^\circ$; Figure 5-7b). This indicates that tectonic-induced scattering or remagnetization can be ruled out. The plot of elongation versus inclination for these data (dashed) and for the TK03.GAD model (solid) is shown in Figure 5-12a. The crossing points represent the inclination/elongation pair most consistent with the TK03.GAD. Consequently, the flattening factor f ($\tan I_c = f \tan I_o$, where I_c and I_o are the compacted and original inclinations; Jackson et al., 1991) is obtained as 0.75 and the mean inclination is corrected from $-39.7^\circ \pm 5.0^\circ$ to -47.3° with 95% confidence limits between -40.0° to -56.1° (Figures 5-12b-c).

Mean ChRMs from two sections are significantly different in the Daheyan area (Figure 5-7c). For the section DHY-A, the PSV is underrepresented within site and among sites ($A95 = 2.6^\circ < A95 \text{ min} = 3.3^\circ$; Figure 5-7e; Table 5-1). Specimens from site TH18 were collected from a ~ 3 m-thick basalt flow (Figure 5-3b), and their site-mean direction can be therefore regarded as a spot record of paleo-geomagnetic field. Considering that basalt flows and sandstones are interbedded in the section DHY-A (Figure 5-3b), remagnetization of the sandstone sites TH19 and TH20 caused by the overlying basalts could have resulted in their underrepresentation of the PSV. This assumption is also supported by the straightforward demagnetization curves of the corresponding specimens (Figures 5-6a, b and f). Hereby, their site-mean directions should be also regarded as spot records of the paleo-geomagnetic field. Consequently, we combined the site-mean directions of sedimentary sites TH19 and TH20 with 35 single directions from the section DHY-B, where the PSV is fully represented ($A95 \text{ min} = 2.9^\circ < A95 = 3.8^\circ < A95 \text{ max} = 8.7^\circ$; Figure 5-7f), to perform an E/I correction. The result shows no intersection between the elongation-inclination curve of our data and that of the geomagnetic TK03 field model (Figure 5-12d). Combining with its sub-circularly distributed VGPs, the strata in the DHY area did not experience inclination shallowing (Figures 5-12e-f).

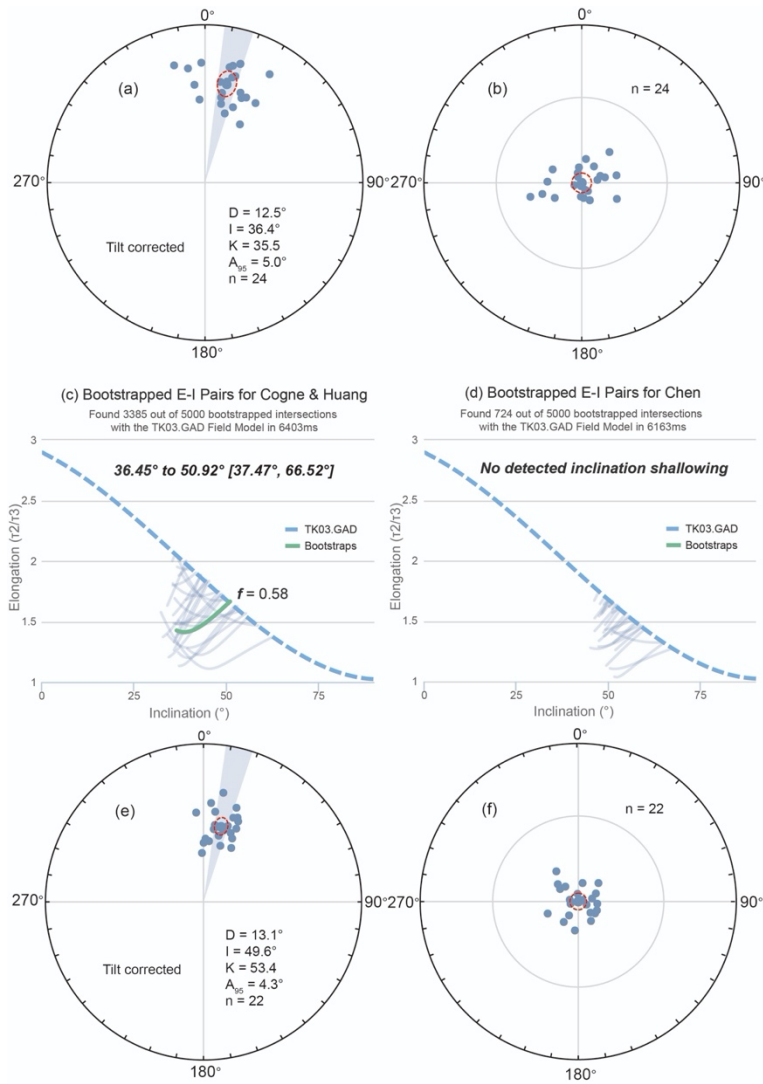


Figure 5-13. Equal-area projections of the (a and e) site-mean magnetic remanent directions and (b and f) corresponding VGPs for the Cretaceous paleomagnetic data for the Turpan-Hami Basin (Cogné et al., 1995; Huang et al., 2004a) and the South Junggar (Chen et al., 1991). (c and d) Tentative E/I correction (Tauxe et al., 2004, 2008) on the Cretaceous paleomagnetic data indicate significant inclination shallowing in the Turpan-Hami Basin (from 36.45° to 50.92° [37.47° , 66.52°]) and no shallowing in the South Junggar.

The anisotropy of magnetic susceptibility or remanence can also be used to correct inclination shallowing, and its accuracy relies on measuring the anisotropy of only ferrimagnetic minerals that carry the ChRMs. In our case, the mean magnetic

susceptibility of the specimens is 1.1×10^{-3} [SI], suggesting that the AMS is controlled by both ferrimagnetic and paramagnetic minerals (Tarling and Hrouda, 1993). Therefore, any correction based on AMS is unsuitable for this study.

5.5.2 Magnetic remanence age and paleomagnetic pole calculation

Due to the limited variation in bedding attitude in the studied strata (Table 5-1), inconclusive fold tests *sensu stricto* were acquired using site-mean directions (McElhinny, 1964). Therefore, a generalized non-parametric fold test was carried out on single ChRMs (Tauxe and Watson, 1994), and this yielded a positive result with the maximum clustering reached at 94-109% unfolding (Figure 5-14), suggesting a pre-tilt origin of the remanence. However, the fold test can only constrain a pre-tilt remanence, and the remanence age is still undeterminable as the age of tilting is difficult to date. Thus, further considerations are required to provide better constraints for the remanence age.

Demagnetization results of the hematite-bearing specimens show three kinds of behaviors (Figure 5-6), which most likely resulted from different formation mechanisms of hematite in sedimentary rocks (Butler, 1992; Kodama, 2012). (1) In some red sandstones, specular hematite likely exists as a primary mineral and carries a detrital remanence which was also recorded by detrital magnetite (Figures 5-6e-f). (2) Pigmentary hematite with submicron size is secondary and shows characteristic color of red beds. In the case that the pigmentary hematite formed within $10^3 - 10^4$ years of deposition, its chemical remanence magnetization (CRM) can still be a useful paleomagnetic signal (Kodama, 2012); but the CRM may be slightly different from the ChRM carried by primary detrital hematite. Thus, the slightly curved demagnetization plots between 590 and 690 °C could be the result of superposition of remanences from different phases of hematite (Figure 5-6g). (3) Pigmentary hematite formed much later (e.g., post-depositional weathering/oxidation under low temperature condition) may carry a remanence that substantially differs from a primary one. In addition, because

hematite has a high blocking temperature, its remanence may act as noise to hamper the accurate ChRM isolation (Figures 5-6h-i).

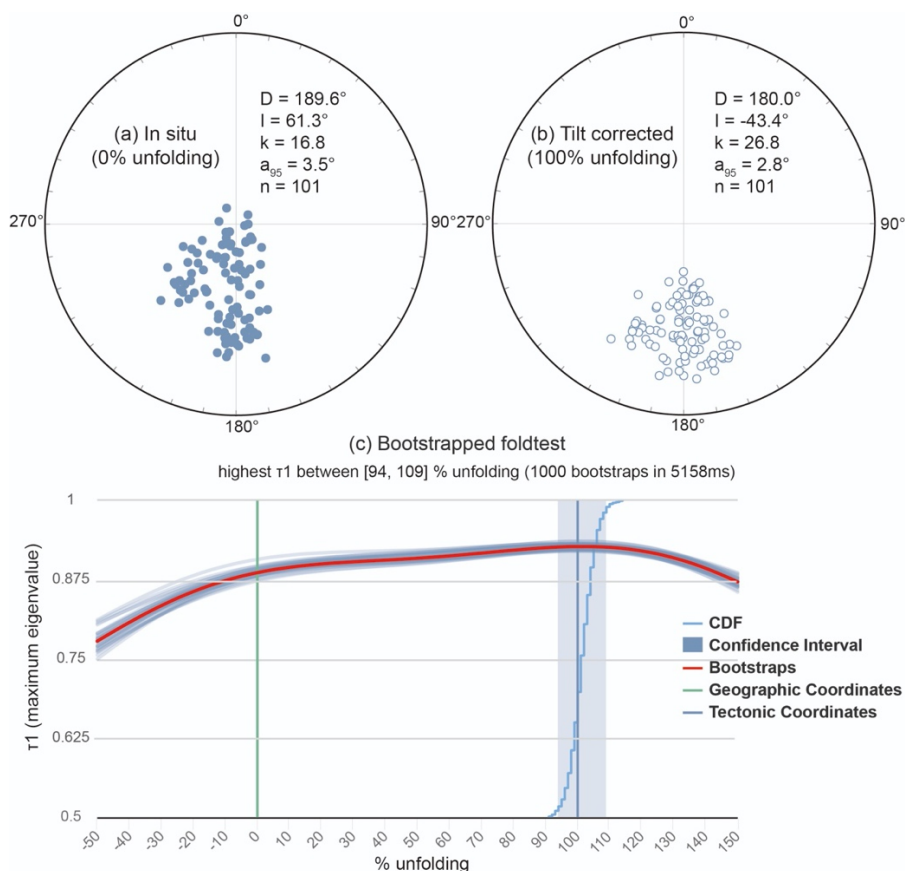


Figure 5-14. Results of non-parametric fold test (Tauxe and Waston, 1994) on the samples from the Qijiaojing and Daheyan areas, northern margin of the Turpan-Hami Basin. Equal-area projections of the individual ChRM directions from each specimen in (a) geographic and (b) stratigraphic coordinates. (c) Results of the fold test with bootstrapped statistics on the first eigenvalues (τ_1 ; Tauxe and Waston, 1994) upon progressive untilting. The first 25 bootstraps are shown. The 95% bootstrap error interval is indicated by the shadow zone, according to which the best grouping is reached at 94-109% untilting.

These demagnetization behaviors indicate that the isolated ChRMs we report should be primary ones. Moreover, the ChRMs may be acquired during or slightly later ($10^3 - 10^4$ years) than the time of deposition. Several lines of evidence, including low

anisotropy of magnetic susceptibility, fully representation of the PSV by ChRMs (except the section DHY-A), significant inclination shallowing of the strata in the Qijiaojing area, and sole reverse polarity of ChRMs (consistent with the Kiaman Reversed Superchron; Belica et al., 2017), all support a primary origin, i.e., the remanence age should be approximately the same as the formation age of the studied strata.

In the Qijiaojing area, the rhyolitic cobble sample 15T123B (310.5 ± 2.4 Ma) and sandstone sample 15T124 (maximum depositional age: 293.9 ± 2.0 Ma) provide an oldest formation age for the Kulai Formation at ~ 294 Ma (Figure 5-11). Together with regional stratigraphic data (XBGMR, 1960; XGSC, 2007), the formation age of the strata can be reasonably constrained as between 294 and 273 Ma. Combined with the Early Permian deposition age of the studied strata in the Daheyan area (297 - 273 Ma; Wali et al., 2018), the magnetic remanence age of our specimens is determined to be Early Permian.

To estimate a high-quality and reliable paleomagnetic pole, three factors were taken into consideration. First, the inclination-corrected ChRMs of specimens from the Qijiaojing area were used in the calculation in order to minimize effects of inclination shallowing. Second, as discussed above, the ChRMs of the specimens from the section DHY-A only provide spot records of the paleo-geomagnetic field and is incomplete instead of the centered axial dipole, therefore their mean direction significantly deviates from those obtained from other sites of coeval strata in the Daheyan and Qijiaojing sections (Figures 5-7a, c and 5-15b). In addition, the distribution of ChRMs from the Daheyan area (Figure 5-7c) possibly suggests syn-depositional local tectonic movements as indicated by the thick conglomerate layers between the two sections (Figure 5-4i). Thus, the results from the DHY-A section are excluded from the pole calculation. Third, because the site-mean VGPs are more circularly distributed than the site-mean directions (Tauxe and Kent, 2004; Deenen et al., 2011), paleomagnetic poles were calculated using Fisher statistics (Fisher, 1953) on the site-mean VGPs (Figures

5-15c-d; Table 5-4) which were derived from the site-mean directions (Figures 5-15a-b; Table 5-4). Consequently, the first Early Permian paleomagnetic pole for the Turpan-Hami Block was obtained as: 75.7°N , 276.3°E , $A_{95} = 5.7^{\circ}$, $K = 72.8$, and $N = 10$ sites (Figure 5-15d; Table 5-4).

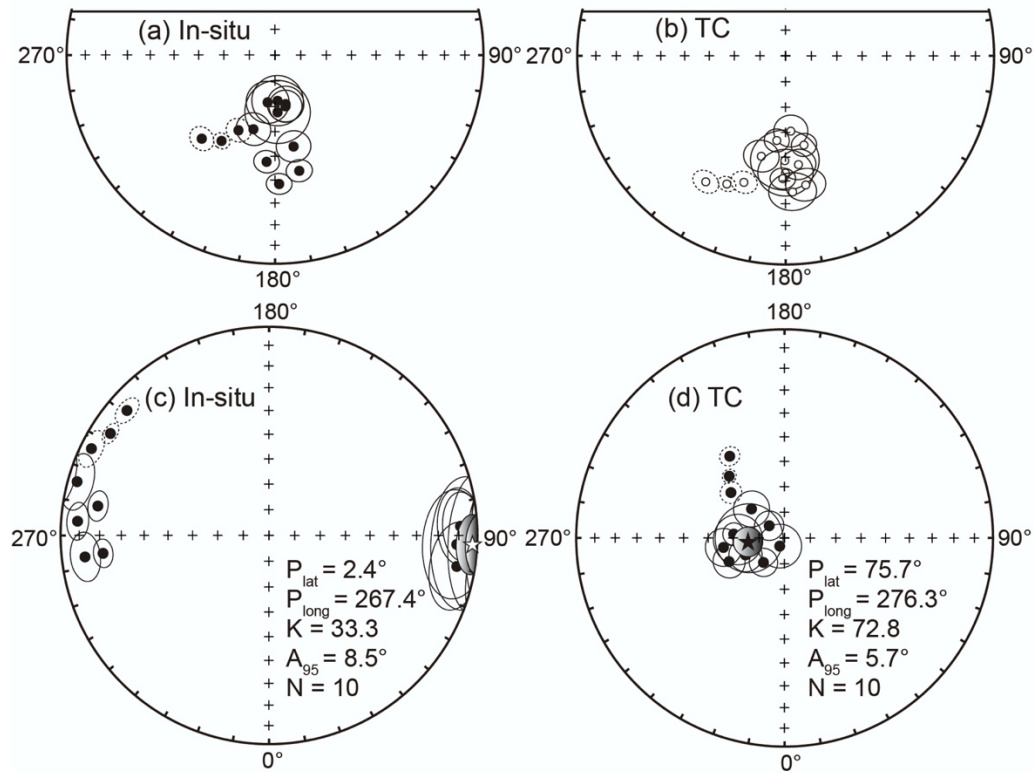


Figure 5-15. Equal-area projections of site-mean magnetic remanent directions in (a) geographic and (b) stratigraphic coordinates; and corresponding visual geomagnetic poles (VGPs) in (c) geographic and (d) stratigraphic coordinates of the Early Permian paleomagnetic samples from the Qijiaojing and Daheyan areas, northern margin of the Turpan-Hami Basin. Noting that the dashed circles stand for Sites H18-H20 which were sampled from the QJJ-A section.

5.6 Summary

(1) Rock magnetic mineralogic analyses, including thermomagnetic experiments, hysteresis measurements, and isothermal remanent magnetization (IRM) acquisition, revealed that the studied samples can be divided into three groups according to their magnetic remanence carrier(s): (Group A) hematite with minor magnetite; (Group B)

magnetite with minor hematite; and (Goup C) magnetite.

(2) After tilt correction, the studied samples show a typical magnetic fabric related to depositional process without intensive later deformation.

(3) Two geochronological samples, one rhyolitic cobble and one sandstone, were collected from the Qijiaojing area. According to field observation, the sandstone was taken from the lower part. The zircon U-Pb LA-ICP-MS dating showed that the rhyolitic cobble sample has a weighted mean age at 310.5 ± 1.2 Ma, and that the maximum depositional age of the sandstone is around 294 Ma.

(4) Based on TK03.GAD model, significant magnetic inclination shallowing with a flattening factor $f = 0.75$ was found for the strata in the Qijiaojing area; in contrast, the strata in the Daheyuan area did not experience magnetic inclination shallowing.

(5) Several lines of evidence, including low anisotropy of magnetic susceptibility, fully representation of the PSV by ChRMs, significant inclination shallowing of the strata in the Qijiaojing area, and sole reverse polarity of ChRMs, all support a primary origin. Applying E/I correction, the first Early Permian paleomagnetic pole for the Turpan-Hami Block was estimated as: 75.7°N , 276.3°E , $A95 = 5.7^\circ$, $K = 72.8$, and $N = 10$ sites.

Table 5-1. Sampling and demagnetization results of Early Permian paleomagnetic samples from Turpan-Hami basin (Fisher statistics on the VGPs correspondent to the ChRMs is used)

Section	Site	Lithology	Slat (N°)	Slong (E°)	Strike (°)	Dip (°)	n/n ₀	Coordinates	mDec (°)	mInc (°)	k	α ₉₅ (°)	K	A ₉₅ (°)	A _{95 min} (°)	A _{95 max} (°)	S (°)	ΔD	ΔI	λ	
Qijiaojing area																					
QJJ-A	TH01	F S	43°21'55.0"	91°40'51.8"	254	62	13/14	Geo.	217.2	62.0	33.2	7.3	14.8	11.1	4.3	16.3	21.3	15.4	9.3	43.2	
								Strati.	193.4	-41.1	33.2	7.3	25.1	8.4	4.3	16.3	16.3	9.2	11.4	-23.6	
	TH02	F/C S			264	64	9/10	Geo.	200.1	78.8	26.9	10.1	9.5	17.6	5.0	20.5	26.5	55.2	9.8	68.4	
								Strati.	180.0	-35.9	26.9	10.1	31.4	9.3	5.0	20.5	14.5	9.9	13.8	-19.9	
	TH03	C S			258	63	8/8	Geo.	187.6	77.7	67.6	6.8	21.6	12.2	5.2	22.1	17.5	32.0	6.9	66.5	
								Strati.	173.2	-38.5	67.6	6.8	60.0	7.2	5.2	22.1	10.5	7.8	10.2	-21.7	
	Mean						30/32	Geo.	208.8	71.6	26.0	5.3	10.5	8.5	3.1	9.6	25.3	15.5	5.5	56.4	
								Strati.	183.8	-39.2	29.6	4.9	25.8	5.3	3.1	9.6	16.0	5.7	7.4	-22.2	
	QJJ-B	TH11	F S	43°21'58.3"	91°41'33.7"	262	63	6/12	Geo.	196.2	74.4	29.6	12.5	11.2	21.0	5.9	26.5	24.6	47.2	12.8	60.8
									Strati.	180.4	-41.0	29.6	12.5	38.2	11.0	5.9	26.5	13.2	12.0	14.9	-23.5
TH12		F S			255	65	12/12	Geo.	178.5	60.4	37.0	7.2	21.8	9.5	4.4	17.1	17.5	12.7	8.2	41.4	
								Strati.	176.1	-53.4	37.0	7.2	31.2	7.9	4.4	17.1	14.6	9.5	8.2	-33.9	
TH13		F S			253	70	7/11	Geo.	217.9	75.1	55.4	8.2	18.5	14.4	5.5	24.1	18.9	32.0	8.6	62.0	
								Strati.	176.8	-28.0	55.4	8.2	62.7	7.7	5.5	24.1	10.3	8.0	12.8	-14.9	
TH14		F S	43°22'00.4"	91°41'26.4"	256	71	8/13	Geo.	187.3	78.0	58.1	7.3	19.2	13.0	5.2	22.1	18.6	35.1	7.3	67.0	
								Strati.	171.0	-30.1	58.1	7.3	56.8	7.4	5.2	22.1	10.8	7.7	12.0	-16.2	
Mean							33/48	Geo.	188.7	70.9	29.4	4.7	13.1	7.2	3.0	9.1	22.7	12.7	4.7	55.3	
								Strati.	175.6	-40.1	24.4	5.2	28.4	4.8	3.0	9.1	15.3	5.2	6.6	-22.8	
QJJ	Mean						63/80	Geo.	198.1	71.5	26.9	3.5	11.5	5.5	2.3	6.0	24.2	10.0	3.6	56.2	
								Strati.	179.6	-39.7	26.0	3.6	26.0	3.6	2.3	6.0	16.0	3.9	5.0	-22.6	
Daheyan area																					
DHY-B	TH15	C S	43°19'15.2"	88°53'15.3"	294	74	11/11	Geo.	189.3	47.1	97.5	4.6	80.3	5.1	4.6	18.1	9.1	5.8	6.1	28.3	
								Strati.	185.8	-56.6	97.5	4.6	56.2	6.1	4.6	18.1	10.9	7.7	5.9	-37.1	

DHY-A	TH16	F S			280	79	12/12	Geo.	172.5	43.5	87.6	4.7	83.1	4.8	4.4	17.1	8.9	5.3	6.2	25.4
								Strati.	168.2	-54.1	87.6	4.7	96.0	4.5	4.4	17.1	8.3	5.4	4.5	-34.6
	TH17	C S			117	85	12/12	Geo.	182.2	38.3	94.6	4.5	101.5	4.3	4.4	17.1	8.1	4.7	6.2	21.6
								Strati.	181.2	-40.9	94.6	4.5	93.5	4.5	4.4	17.1	8.4	4.9	6.1	-23.4
	Mean						35/35	Geo.	181.0	43.1	62.9	3.1	55.1	3.3	2.9	8.7	11.0	3.6	4.3	25.1
								Strati.	178.3	-50.6	48.4	3.5	42.5	3.8	2.9	8.7	12.5	4.4	4.2	-31.3
	TH18	Basalts	43°18'26.5"	88°53'45.5"	258	82	7/7	Geo.	224.6	43.3	169.8	4.6	169.0	4.7	5.5	24.1	6.2	5.1	6.0	25.2
								Strati.	212.3	-29.5	169.8	4.6	237.7	3.9	5.5	24.1	5.3	4.1	6.4	-15.8
	TH19	F S			259	82	12/14	Geo.	216.1	47.9	189.3	3.2	148.3	3.6	4.4	17.1	6.7	4.1	4.2	29.0
								Strati.	204.5	-33.1	189.3	3.2	274.8	2.6	4.4	17.1	4.9	2.8	4.1	-18.1
	TH20	F S			259	78	6/8	Geo.	211.3	55.4	204.4	4.7	167.5	5.2	5.9	26.5	6.3	6.4	5.1	36.0
								Strati.	198.1	-36.5	204.4	4.7	171.9	5.1	5.9	26.5	6.2	5.5	7.5	-20.3
	Mean						25/29	Geo.	217.7	48.5	103.7	2.9	88.8	3.1	3.3	10.8	8.6	3.6	3.6	29.5
								Strati.	205.3	-33.0	113.3	2.7	123.7	2.6	3.3	10.8	7.3	2.7	4.1	-18.0

Abbreviations: F S, C S stand for fine-grained sandstones and coarse-grained sandstones; Slat (N°)/ Slong (E°), latitude and longitude of sampling sites; n/n₀, number of specimens in calculation/ totally studied.

mDec, mInc, k and α_{95} stand for mean declination, mean inclination, precision parameter and 95% confidence interval of directions.

K, A₉₅ and S correspond to precision parameter, 95% confidence interval and standard angular deviation of Visual Geomagnetic Poles (VGPs).

A₉₅ max, A₉₅ min mean the highest and lowest 95% confidence interval which can present the paleosecular variation of the geomagnetic field.

ΔD , ΔI and λ signify error of declination, error of inclination and paleolatitude.

These calculations were performed using *paleomagnetism.org* (Koymans et al., 2016) which is based on Deenen et al. (2011) and Fisher (1953).

Table 5-2. Zircon LA-ICP-MS U-Pb analytical results of the geochronological samples 15T123B and 15T124

Analysis <i>no.</i>	Corrected ratios						Corrected ages (Ma)						Error correlation	Comments
	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ	²⁰⁷ Pb/ ²³⁵ U	1σ	²⁰⁶ Pb/ ²³⁸ U	1σ	²⁰⁷ Pb/ ²⁰⁶ Pb	1σ	²⁰⁷ Pb/ ²³⁵ U	1σ	²⁰⁶ Pb/ ²³⁸ U	1σ		
15T123B Rhyolitic gravel														
15T123B-01	0.05251	0.0039	0.35793	0.0255	0.04944	0.001	308	170	311	19	311	6	0.02	Concordant
15T123B-02	0.05313	0.002	0.35992	0.0126	0.04914	0.0006	334	86	312	9	309	4	0.02	Concordant
15T123B-03	0.05231	0.0031	0.35597	0.0199	0.04938	0.001	299	137	309	15	311	6	0.02	Concordant
15T123B-04	0.05228	0.0018	0.35693	0.0116	0.04953	0.0006	298	80	310	9	312	4	0.02	Concordant
15T123B-05	0.05298	0.0037	0.35745	0.0241	0.04897	0.001	328	162	310	18	308	6	0.01	Concordant
15T123B-06	0.05266	0.0023	0.35639	0.0146	0.0491	0.0007	314	101	310	11	309	4	0.01	Concordant
15T123B-07	0.05235	0.0029	0.35994	0.0187	0.0499	0.0009	301	128	312	14	314	6	0.03	Concordant
15T123B-08	0.05262	0.0021	0.35825	0.0138	0.04939	0.0007	312	95	311	10	311	4	0.03	Concordant
15T123B-09	0.05266	0.0035	0.3596	0.0228	0.04954	0.0011	314	154	312	17	312	7	0.04	Concordant
15T123B-10	0.05281	0.0024	0.36047	0.0154	0.04952	0.0007	321	105	313	12	312	5	0.03	Concordant
15T123B-11	0.05322	0.002	0.3578	0.0128	0.04879	0.0007	338	88	311	10	307	4	0.02	Concordant
15T123B-12	0.05298	0.0029	0.37956	0.0199	0.05198	0.0009	328	128	327	15	327	5	0.03	Concordant
15T123B-13	0.05232	0.0017	0.35546	0.0111	0.0493	0.0006	299	77	309	8	310	4	0.04	Concordant
15T123B-14	0.05219	0.0028	0.35544	0.0184	0.04942	0.0007	294	125	309	14	311	4	0.02	Concordant
15T123B-15	0.05232	0.004	0.3599	0.0263	0.0499	0.0012	299	175	312	20	314	7	0.04	Concordant
15T123B-16	0.05213	0.0051	0.35706	0.0338	0.04968	0.0015	291	226	310	25	313	9	0.03	Concordant
15T123B-17	0.05213	0.0057	0.3571	0.038	0.04969	0.0015	291	251	310	28	313	9	0.02	Concordant
15T124 Sandstones														
15T124-01	0.05239	0.0028	0.3545	0.0179	0.04907	0.0008	302	123	308	13	309	5	0.03	Concordant
15T124-02	0.05304	0.0019	0.35889	0.012	0.04908	0.0006	331	82	311	9	309	4	0.03	Concordant
15T124-03	0.05255	0.0023	0.35551	0.0146	0.04907	0.0007	309	101	309	11	309	4	0.03	Concordant
15T124-04	0.05204	0.0028	0.33368	0.0168	0.04649	0.0008	287	124	292	13	293	5	0.01	Concordant
15T124-05	0.05308	0.0016	0.35677	0.0102	0.04876	0.0006	332	71	310	8	307	3	0.02	Concordant
15T124-06	0.05364	0.0035	0.37878	0.0242	0.05123	0.0009	356	153	326	18	322	6	0.01	Concordant
15T124-07	0.0522	0.0018	0.33418	0.0111	0.04644	0.0006	294	82	293	8	293	3	0.03	Concordant

15T124-08	0.05256	0.0024	0.35514	0.0153	0.04901	0.0007	310	105	309	11	308	4	0.02	Concordant
15T124-09	0.05411	0.0046	0.44672	0.0367	0.05988	0.0012	376	194	375	26	375	7	0.01	Concordant
15T124-10	0.05259	0.0022	0.3513	0.0141	0.04846	0.0007	311	98	306	11	305	4	0.02	Concordant
15T124-11	0.05252	0.0024	0.34962	0.0156	0.04829	0.0007	308	109	304	12	304	4	0.02	Concordant
15T124-12	0.05252	0.0031	0.35549	0.0204	0.0491	0.0009	308	139	309	15	309	5	0.02	Concordant
15T124-13	0.05292	0.0056	0.37834	0.0388	0.05185	0.0015	325	243	326	29	326	9	0.01	Concordant
15T124-14	0.05612	0.004	0.55456	0.0377	0.07169	0.0015	457	161	448	25	446	9	0.02	Concordant
15T124-15	0.05249	0.0025	0.35164	0.016	0.04859	0.0008	307	112	306	12	306	5	0.02	Concordant
15T124-16	0.05271	0.0018	0.35807	0.0119	0.04927	0.0006	316	81	311	9	310	4	0.03	Concordant
15T124-17	0.05242	0.0027	0.3482	0.0175	0.04818	0.0007	304	122	303	13	303	4	0.02	Concordant
15T124-18	0.05256	0.0024	0.35135	0.0153	0.04849	0.0008	310	107	306	12	305	5	0.03	Concordant
15T124-19	0.05258	0.0018	0.34821	0.0113	0.04803	0.0006	311	80	303	8	302	4	0.02	Concordant
15T124-20	0.05263	0.0023	0.35257	0.0149	0.04859	0.0007	313	103	307	11	306	4	0.01	Concordant
15T124-21	0.0524	0.0017	0.3523	0.011	0.04877	0.0005	303	77	306	8	307	3	0.02	Concordant
15T124-22	0.0561	0.0027	0.56655	0.0262	0.07325	0.0011	456	110	456	17	456	6	0.03	Concordant
15T124-23	0.05241	0.0028	0.3536	0.0181	0.04893	0.0007	303	123	307	14	308	4	0.02	Concordant
15T124-24	0.06468	0.0022	1.1116	0.0349	0.12464	0.0016	764	72	759	17	757	9	0.02	Concordant
15T124-25	0.05257	0.0015	0.35578	0.0098	0.04909	0.0006	310	68	309	7	309	3	0.05	Concordant
15T124-26	0.05263	0.0017	0.35647	0.0107	0.04913	0.0006	313	74	310	8	309	3	0.02	Concordant
15T124-27	0.05325	0.0015	0.39942	0.0106	0.0544	0.0006	339	66	341	8	341	4	0.01	Concordant
15T124-28	0.05235	0.0012	0.34564	0.0071	0.04789	0.0005	301	52	301	5	302	3	0.06	Concordant
15T124-29	0.05632	0.0017	0.57857	0.0167	0.0745	0.0008	465	70	464	11	463	5	0.03	Concordant
15T124-30	0.05243	0.002	0.347	0.0128	0.048	0.0006	304	90	302	10	302	4	0.02	Concordant
15T124-31	0.05261	0.0018	0.35913	0.0115	0.04951	0.0006	312	79	312	9	312	4	0.01	Concordant
15T124-32	0.05249	0.0035	0.35314	0.0228	0.04879	0.0008	307	153	307	17	307	5	0.01	Concordant
15T124-33	0.05287	0.0012	0.37171	0.0079	0.05099	0.0005	323	54	321	6	321	3	0.01	Concordant
15T124-34	0.05281	0.0022	0.34638	0.0135	0.04758	0.0007	321	96	302	10	300	4	0.01	Concordant
15T124-35	0.05201	0.0026	0.33389	0.0159	0.04655	0.0007	286	116	293	12	293	4	0.01	Concordant
15T124-36	0.05275	0.0026	0.35598	0.0168	0.04895	0.0008	318	115	309	13	308	5	0.01	Concordant
15T124-37	0.05343	0.0034	0.40153	0.0247	0.05448	0.0009	347	148	343	18	342	6	0.01	Concordant

15T124-38	0.05291	0.0037	0.35274	0.0241	0.04836	0.0008	325	162	307	18	304	5	0.01	Concordant
15T124-39	0.05321	0.0019	0.40068	0.0134	0.05464	0.0007	338	82	342	10	343	4	0.01	Concordant
15T124-40	0.05277	0.0033	0.36704	0.0222	0.05045	0.0008	319	146	317	17	317	5	0.01	Concordant
15T124-41	0.05593	0.0019	0.55776	0.0175	0.07233	0.0009	450	76	450	11	450	5	0.01	Concordant
15T124-42	0.05254	0.0023	0.35905	0.0151	0.04954	0.0007	309	103	312	11	312	4	0.01	Concordant
15T124-43	0.05453	0.0045	0.44653	0.0356	0.05937	0.0013	393	189	375	25	372	8	0.01	Concordant
15T124-44	0.05345	0.0037	0.40016	0.0263	0.05431	0.0012	348	159	342	19	341	7	0.01	Concordant
15T124-45	0.06593	0.0043	1.21132	0.0754	0.13328	0.0028	804	140	806	35	807	16	0.01	Concordant
15T124-46	0.05234	0.0031	0.34326	0.0197	0.04754	0.0008	300	139	300	15	299	5	0.01	Concordant
15T124-47	0.05257	0.0013	0.36023	0.0085	0.0497	0.0005	310	59	312	6	313	3	0.01	Concordant
15T124-48	0.05272	0.0015	0.36436	0.01	0.05013	0.0006	317	68	315	7	315	3	0.01	Concordant
15T124-49	0.05823	0.0026	0.69369	0.0296	0.08642	0.0013	538	100	535	18	534	8	0.01	Concordant
15T124-50	0.07809	0.0011	2.09283	0.024	0.19438	0.0016	1149	27	1146	8	1145	9	0.01	Concordant

Table 5-3. Compilation of Cretaceous Paleomagnetic results in the Turpan-Hami and Junggar blocks and tentative E/I corrections

Site	n/N	Dg (°)	Ig (°)	Ds (°)	Is (°)	k	a ₉₅ (°)	Age	E/I correction (<i>f</i> = 0.58)	
									Ds-c (°)	Is-c (°)
Cogne, 1995, EPSL (Shengjinkou 43° N, 90° E)										
2R	4/10	229.1	-52.3	196.5	-22.7	180.0	6.9	Late Cretaceous - Early Tertiary (K ₂ -E ₁)	196.5	-35.7
4	6/6	87.3	58.5	11.5	47.8	132.0	6.0		11.5	62.2
5	3/4	258.9	-43.1	212.4	-40.1	222.0	9.7		212.4	-55.4
7	4/7	68.0	-68.4	176.2	-46.2	27.0	18.0	Early Cretaceous (K ₁)	176.2	-60.9
10	4/6	220.2	-75.9	210.8	-21.2	48.0	13.4		210.8	-33.7
13	6/6	170.5	-57.0	194.5	-31.7	14.0	18.6		194.5	-46.7
14	9/9	241.8	7.2	246.1	37.0					
15	5/6	116.8	18.7	109.0	17.1					
16	4/6	358.7	76.6	25.8	39.9	54.0	12.6		25.8	55.2
18	4/6	134.6	-71.7	190.7	-44.3	54.0	14.7		190.7	-59.2
19	7/7	29.4	6.8	15.9	52.4	67.0	7.4		15.9	65.9
11	5/5	168.1	74.2	30.9	54.6	44.0	12.4	Late Jurassic (J ₃)	30.9	67.6
12	7/7	358.7	63.6	21.5	38.7	80.0	8.9		21.5	54.0
Huang et al. 2004, Science in China (Shengjinkou 43.93° N 89.58° E)										
XTK10	6/8	183.0	-54.8	193.4	-24.9	88.2	9.3	Early Cretaceous (K ₁)	193.4	-38.6
XTK11	8/8	164.1	-49.3	178.2	-25.0	70.5	6.6		178.2	-38.7
XTK12	6/8	335.8	57.9	351.5	27.7	69.6	10.5		351.5	42.1
XTK13	8/9	173.0	-73.6	200.3	-47.8	21.9	12.1		200.3	-62.2
XTK14	8/9	148.3	-27.4	165.2	-24.2	135.0	5.6		165.2	-37.7
XTK15	5/6	173.9	-59.5	188.7	-35.7	1227.0	3.2		188.7	-51.0
XTK16	9/9	9.2	72.3	23.3	41.3	23.8	10.8		23.3	56.5
XTK38	6/6	174.4	-57.1	196.6	-22.5	121.6	7.9		196.6	-35.5
XTK41	6/7	254.9	63.8	12.3	34.8	28.3	12.8		12.3	50.1
XTK42	4/5	355.9	50.5	12.0	26.2	105.1	9.0		12.0	40.2
Tuyugou 42.88° N 89.70° E										

XTK19	6/8	67.3	60.8	15.9	30.6	22.1	14.6		15.9	45.5
XTK21	7/8	242.9	-79.2	190.7	-41.7	62.0	7.7	Early Cretaceous (K ₁)	190.7	-56.9
XTK23	4/5	263.9	-79.2	173.8	-37.7	130.0	11.0		173.8	-53.0
Mean	24			12.5	36.4	32.3	5.3		12.3	50.9
Chen et al., 1991, JGR (Manas 44.2° N 86.0° E)										
1	7/9	171.8	77.6	21.5	51.3	15.1	17.8			
2	6/11	33.7	76.4	19.5	42.9	9.9	22.4			
4	7/10	23.2	73.3	10.3	30.5	49.9	8.6			
19	8/8	310.1	-42	181.8	-57.1	51.1	7.8			
20	8/9	305.7	-36.6	196.7	-59.6	108.8	5.3			
21	8/8	313.2	-34.4	180.1	-59.3	50.8	8.6			
25	7/8	196.7	20.1	202	-42	36.7	10.1	Late Jurassic - Early		
28	8/9	208.5	-22.6	204.1	-47.9	18.2	13.3	Cretaceous (J ₃ -K ₁)		
29	6/11	343.6	79.3	355.3	42.7	97.9	9.3			
30	9/11	348.1	66.1	8	51.6	118.6	5.1			
31	7/8	50.8	77.7	24.3	53.7	25.1	18.7			
32	8/8	36.9	66.1	23.3	43.8	57.5	8			
33	6/8	3.6	68.9	7.4	42	34.8	13.1			
5	5/9	188.7	1.3	195.6	-51.2	19.9	17.6			
10	7/9	42.2	85.3	4.4	37.3	22.8	12.9			
16	6/6	126.5	49.1	8.4	49.3	191	4.9			
17	7/8	320.8	-29.1	178.2	-64.7	51.3	8.5			
18	8/8	106.2	46.2	19.3	37	28.6	10.5			
26	2/7	2.3	-5.6	5.2	58.3	-		Late Cretaceous - Early		
34	4/10	28.9	28.6	27.2	58.5	17.7	22.5	Teriary (K ₂ -E ₁)		
35	4/9	21.9	17	17.8	48.2	9.9	26.9			
36	2/5	11.4	22.2	12.9	54.6	-				
Mean	22			13.1	49.6	58.7	4.1			

Abbreviations: n/n₀, number of specimens (sites) in calculation/ totally studied.

D_g, I_g, D_s and I_s, declination (D) and inclination (I) in geological (g) and stratigraphic (s) coordinates.

k, precision parameter; α_{95} , 95% confidence interval; f stands for flattening factor (Jackson et al., 1991).

Sites 14 and 15 (in yellow) were rejected by the 45° cut-off criterion (Deenen et al., 2011).

Data from Cogne et al. (1995) and Huang et al. (2004) were supposed to be affected by inclination shallowing with a flattening factor of 0.58.

Data from Chen et al. (1991) were found no obvious inclination shallowing effect (Figure 5-13).

Table 5-4. Paleomagnetic results of Early Permian samples from Turpan-Hami basin (inclination shallowing corrected).

Section	Site	n/n ₀ (N)	D _g (°)	I _g (°)	D _s (°)	I _s (°)	k	α ₉₅ (°)	Plat _g (N°)	Plong _g (E°)	A _{95g} (°)	Plat _s (N°)	Plong _s (E°)	A _{95s} (°)
Qijiaojing area														
QJJ-A	TH01	13/14	202.5	58.5	193.4	-49.1	41.5	6.5	4.9	254.4	8.3	72.9	228.8	7.0
	TH02	9/10	188.1	72.0	179.9	-43.3	28.2	9.9	10.6	87.2	16.4	71.8	272.0	9.7
	TH03	8/8	178.4	70.1	173.1	-46.5	78.1	6.3	7.5	92.6	10.1	73.4	293.6	6.5
QJJ-B	TH11	6/12	186.5	67.6	180.3	-48.5	29.8	12.5	4.1	87.6	19.1	76.1	270.6	13.3
	TH12	12/12	174.1	53.8	175.8	-60.4	47.1	6.4	12.1	276.7	7.5	86.3	329.7	8.5
	TH13	7/11	199.2	70.5	176.9	-35.1	54.6	8.2	9.4	80.6	13.2	65.8	278.9	7.2
	TH14	8/13	178.2	70.9	171.0	-37.6	63.4	7.0	8.7	92.7	11.4	66.4	293.1	6.3
Daheyuan area														
DHY-B	TH15	11/11	189.3	47.2	185.8	-56.5	97.4	4.6	17.8	260.3	4.8	82.4	231.6	5.7
	TH16	12/12	172.5	43.5	168.2	-54.1	87.5	4.7	21.0	276.2	4.6	77.4	319.4	5.5
	TH17	12/12	183.9	57.3	181.2	-40.9	94.3	4.5	8.7	265.8	5.6	70.1	265.7	4.3
DHY-A	TH18	7/7	224.6	43.3	212.3	-29.5	169.7	4.6	10.2	228.7	4.5	51.1	213.9	3.8
	TH19	12/14	216.1	47.9	204.5	-33.1	189.7	3.2	10.5	237.3	3.4	57.4	221.9	2.7
	TH20	6/8	211.2	55.4	198.1	-36.5	204.8	4.7	5.8	244.0	5.7	62.5	229.8	4.2
Mean³		3/13	217.8	49.0			115.1	11.5	8.9	236.7	12.3			
					205.2	-33.2	137.6	10.6				57.2	221.1	10.9
Mean¹³		13/13	194.2	59.5			31.9	7.5	4.0	260.4	9.7			
					185.6	-44.7	34.3	7.2				73.3	254.7	7.6
Mean¹⁰		10/13	184.5	61.5			48.5	7.0	2.4	267.4	8.5			
					178.4	-47.4	72.1	5.7				75.7	276.3	5.7

Abbreviations: n/n₀ (N), number of specimens (sites) in calculation/ totally studied.

D_g, I_g, D_s and I_s, declination (D) and inclination (I) in geological (g) and stratigraphic (s) coordinates; k, precision parameter; α₉₅, 95% confidence interval.

Plat_g, Plong_g, Plat_s, Plong_s and A₉₅, latitude and longitude of paleomagnetic poles in geographic (g) and stratigraphic (s) coordinates, and 95% confidence interval.

Mean³, the mean calculation of the sites from the DHY-A section (Sites TH18, 19 and 20). **Mean¹³** and **Mean¹⁰**, the mean calculation with and without **Mean³**.

Mean¹³ is statistically equal to **Mean¹⁰** at 95% confidence level, with an angular difference of $5.1^\circ \pm 7.0^\circ$. However, in order to be more accurate, we apply **Mean¹⁰** in further study.

Chapter 6. A Multidisciplinary Study on the Baluntai Migmatite (Chinese Central Tianshan)

As one of the important components, understanding the tectonic setting and paleogeographic position of the Chinese Central Tianshan is crucial for unveiling the tectonic evolution history of the Tianshan Belt and even that of the CAOBS (Figures 6-1a-b; Zhong et al., 2015). However, until now, such studies are rare and no paleomagnetic data were carried out on the Chinese Central Tianshan.

With the geological characteristics, it is easy to understand that although the paleomagnetic data are crucial for understanding the tectonic evolution of the Chinese Central Tianshan, no such data were reported: (1) early Paleozoic volcanic/sedimentary sequences are pervasively metamorphosed and deformed; (2) late Paleozoic sedimentary rocks are rare, only some tightly folded Carboniferous sedimentary rocks. In addition, we have performed paleomagnetic work on the Carboniferous sedimentary rocks and found that no meaningful characteristic remanent magnetizations can be isolated (unpublished data); (3) late Paleozoic undeformed plutons are not ideal targets for paleomagnetic work since no tilt-correction can be made as sedimentary cover is nonexistent.

Nevertheless, the widespread migmatites with stable and consistent foliation in the Baluntai area could be a possible target (Figure 6-1c). Once we understand the geological meaning of the foliation, it can serve as a tilt-correction tool. On the other hand, migmatite itself has important significance since it is of great importance in understanding regional tectonic and geodynamic setting; moreover, the mechanism of migmatization and its emplacement (exhumation) are still in debate.

Consequently, a multidisciplinary study on the Baluntai migmatite, including petrography, geochronology, rock magnetism, magnetic fabric, and paleomagnetism, was carried out, which may shed light on the tectonic evolution of the SW CAOBS.

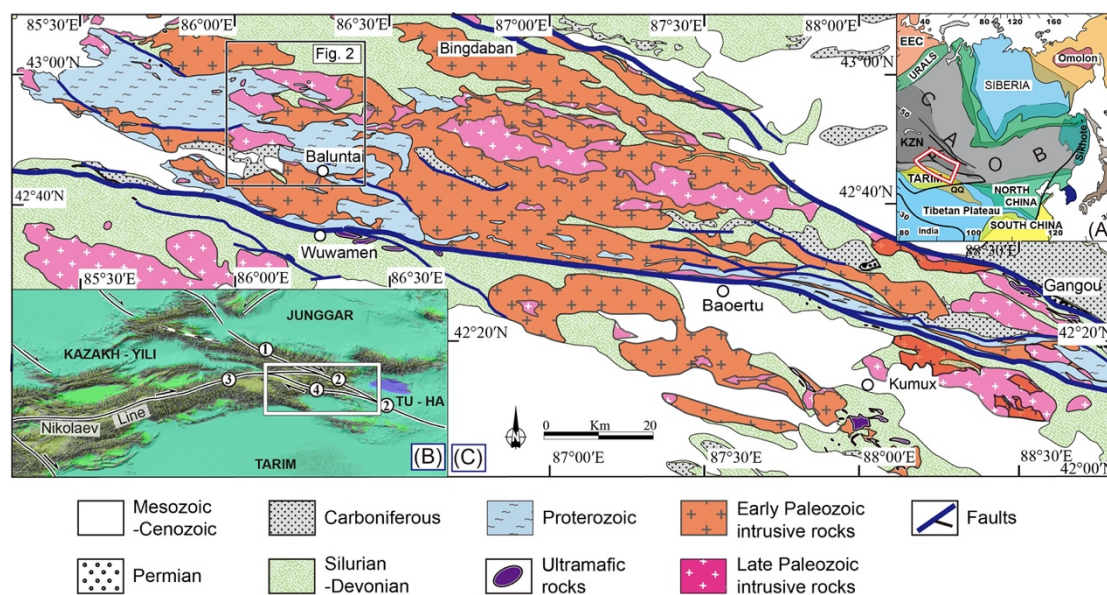


Figure 6-1. (a) Simplified tectonic division of the eastern Asia (after Jahn, 2004). EEC = East European Craton; KZN = Kazakhstan; Q = Qaidam. (b) Tectonic division of the Tianshan Orogenic Belt. Major faults: 1. North Tianshan Fault = NTF; 2. Middle Tianshan Shear Zone = MTSZ; 3. Nalati Fault = NF; and 4. Baluntai Fault = BF. (c) Geological map of the Baluntai Block showing our studied area (Zhong et al., 2015).

6.1 Geological background and sampling

Two fieldworks have been performed on the Baluntai area, and our sampling was conducted in two sections (A and B, Figure 6-2). Section A is located 30 km to the northwest of the Baluntai town. Sediments-derived migmatites crop out in this section with well-developed and consistent foliation whose dip direction and dip angle range from 133° - 149° and 42° - 58° , respectively (Figure 6-3). Thirteen sites with 182 paleomagnetic cores covering leucosomes, mesosomes, and melanosomes were collected from this section. Section B is located 10 km to the northeast of the Baluntai town. Metadiorite-derived migmatites are found in this section with unstable and dispersive foliation whose dip direction and dip angle vary from 161° - 236° and 33° - 81° , respectively. Seven sites with 71 paleomagnetic cores were taken in this section, covering leucosomes, mesosomes, and melanosomes as well.

A migmatitic granite and a meta-greywacke were collected for U-Pb zircon dating

and $^{40}\text{Ar}/^{39}\text{Ar}$ dating. A melanosome was also taken for $^{40}\text{Ar}/^{39}\text{Ar}$ dating and monazite dating.

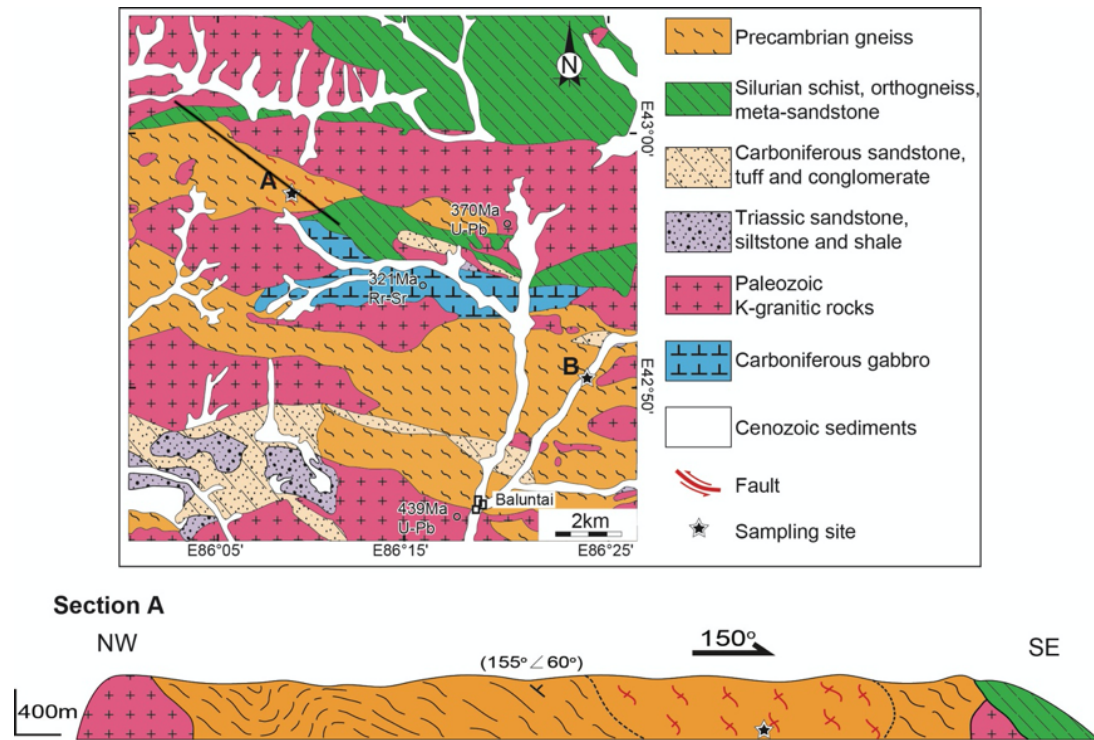


Figure 6-2. Detailed geological map showing our studied locations and a cross-section draw of the section A.

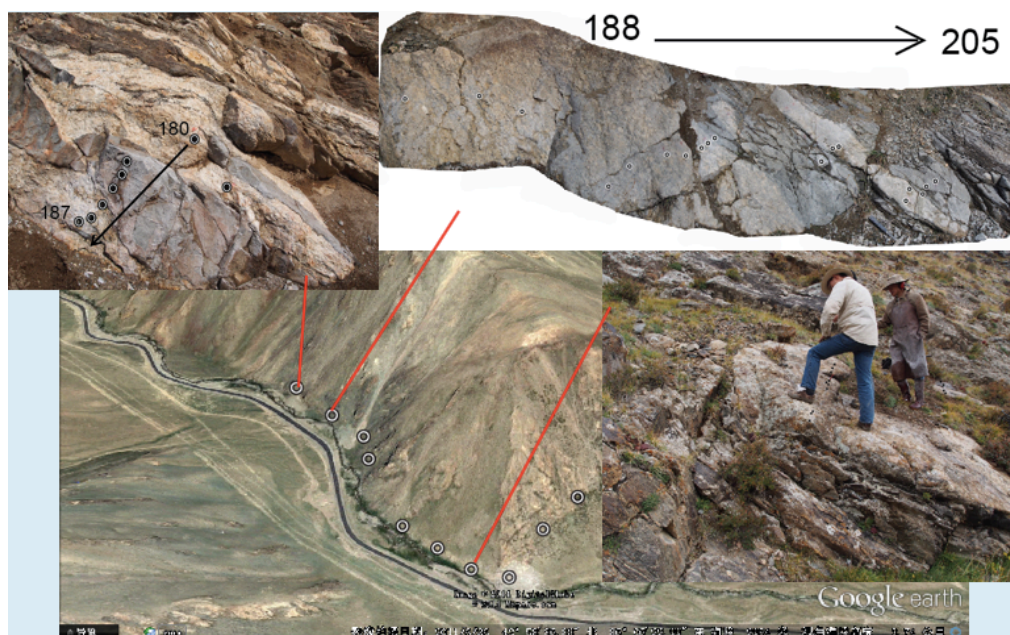


Figure 6-3. Sampling locations and field photos of the migmatites in the Baluntai area. 188-205 stands for sample number.

6.2 Petrographic observations

Representative thin sections of the Baluntai migmatite were observed under a microscope. The leucosomes contain plagioclase (Pl: 35%), quartz (Qtz: 35%), microcline (Mc: 20%), garnet (Grt: 3%), and muscovite (Ms: 1%) together with other minerals (Figures 6-4a-b). It can be easily observed that the quartz and feldspar in the leucosomes are weakly deformed, which is a typical magmatic structure (Figures 6-4a-b). The mesosomes are mainly composed of Qtz (35%), Pl (40%), Bt (20%), and Ms (3%) (Figure 6-4c). The melanosomes are mainly made of Qtz (25%), Pl (40%), and Bt (30%) (Figure 6-4d). A moderate degree of deformation was found in the mesosomes and melanosomes, characterized by the oriented arrangement and rotational structure of biotite, and the elongation, wave extinction, and fine graining of quartz (Figure 6-4c-d).

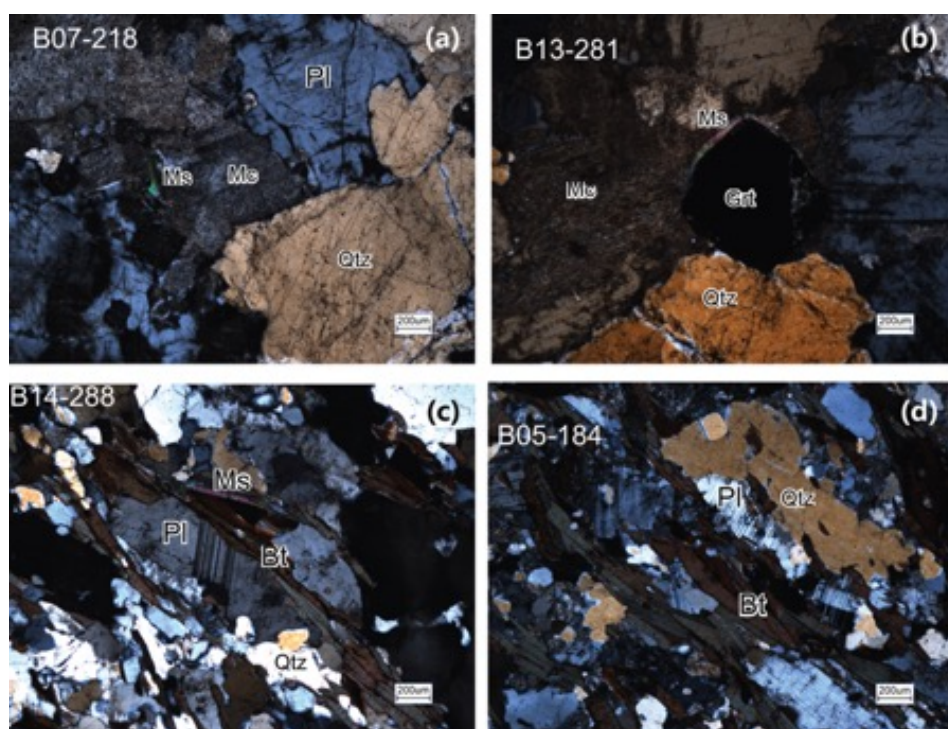


Figure 6-4. Petrographic observations of representative samples: (a-b) leucosomes, (c) mesosomes, (d) melanosomes.

6.3 Multi-minerals thermochronology

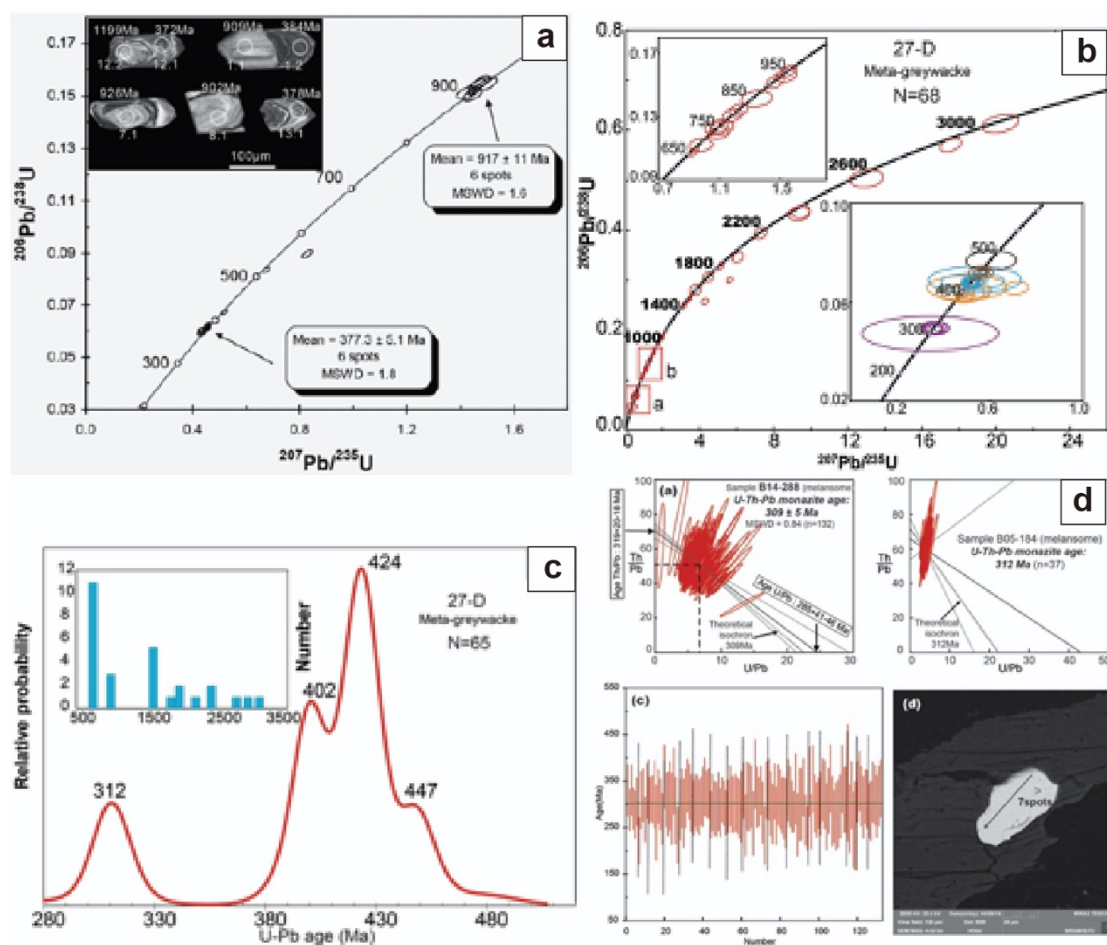


Figure 6-5. (a) LA-ICP-MS U-Pb dating results and CL images of zircons with a core-rim structure collected from migmatitic granite; (b-c) LA-ICP-MS U-Pb dating results and probability density function plot of migmatitic meta-greywacke; (d) U-Th-Pb age of monazites selected from melanosome samples.

For the migmatitic granite sample, the dated zircons are 100-200 μm in size and have subhedral to euhedral shapes with length/width ratios of 1.2-3. Their CL images show clear oscillatory zonings, indicating an origin from a siliceous melt (Figure 6-5a). A typical core-rim structure is developed in these zircons, and they were both dated when possible. A Precambrian age of 917 ± 11 Ma with MSWD = 1.6 ($n = 6$) and a Devonian age of 377.3 ± 5.1 Ma with MSWD = 1.8 ($n = 6$) were obtained. The Precambrian zircons should be inherited zircons from the basement, and the

crystallization age of this migmatitic granite is of Late Devonian.

Detrital zircons from the meta-greywacke sample 27-D are 50-150 μm in size and have sub-rounded to rounded shapes with length/width ratios of 1-2. Their CL images are generally grey or bright without oscillatory zonings. Sixty-eight concordant ages ranging from ~ 299 Ma to ~ 3000 Ma were obtained, which define several peaks (Figures 6-5b-c). Phanerozoic peaks are distributed at ~ 447 , ~ 424 , ~ 402 , and ~ 312 Ma. The youngest peak is ~ 312 Ma, which probably suggests the age of migmatization (Figure 6-5c).

Monazites picked from biotite of two melanosome samples B05-184 and B14-288 give two U-Th-Pb ages at 312 Ma and 309 ± 5 Ma, respectively (Figure 6-5d). Since the blocking temperature of the monazite U-Th-Pb system is close to the melting temperature of siliceous rocks, we propose that the migmatization probably occurred during 309-312 Ma. This proposition is supported by the youngest zircon U-Pb age peak of the meta-graywacke sample.

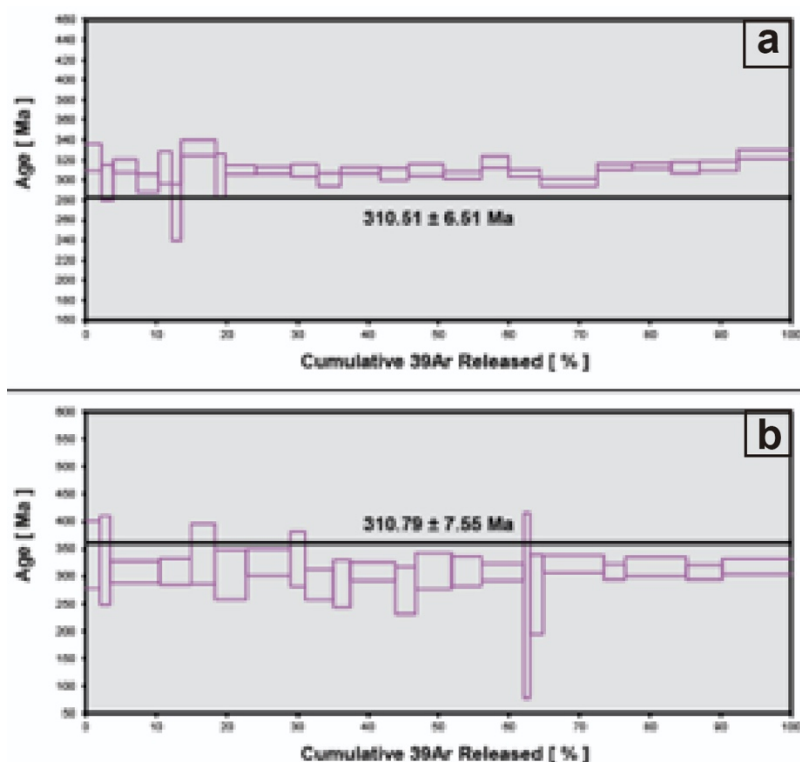


Figure 6-6. (a) muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dating result of the migmatitic granite; (b) biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dating result of the migmatitic meta-greywacke.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating runs were performed on muscovite and biotite selected from migmatitic granite and melanosome. They show plateau ages over all the steps at 310.5 ± 6.5 Ma and 310.8 ± 7.6 Ma, respectively (Figure 6-6)

6.4 Rock magnetism

For leucosomes, hysteresis loops show a mix of paramagnetic and ferromagnetic minerals while for some leucosomes, the diamagnetic minerals are dominant (Figures 6-5 and 6-S1, S2); the thermomagnetic curve indicates that magnetite exists; the IRM acquisition curve supports this conclusion (Figure 6-7a).

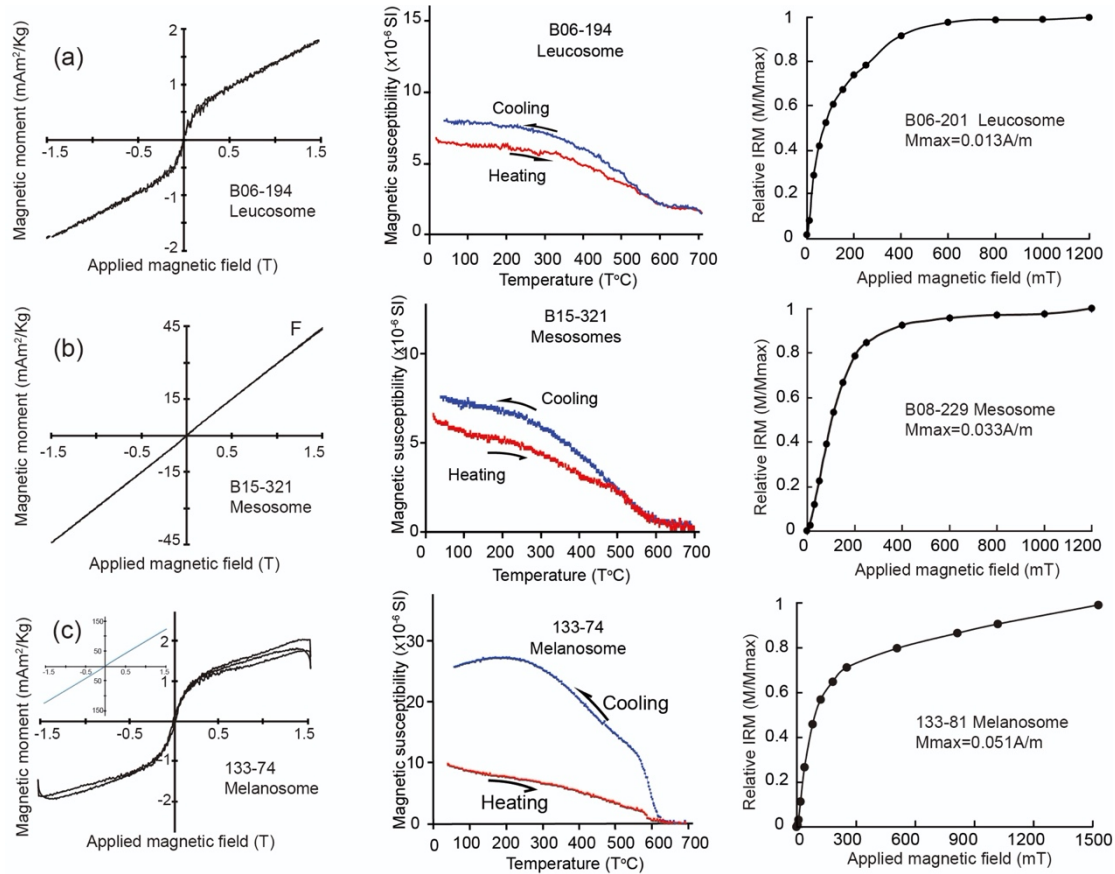


Figure 6-7. Representative results of rock magnetic experiments (room-temperature hysteresis loops, thermomagnetic curves, IRM acquisition curves) of (a) leucosomes; (b) mesosomes; and (c) melanosomes.

For mesosomes, the most important features are the low content of ferromagnetic minerals (magnetite with possible pyrrhotite) and relatively higher content of paramagnetic minerals indicated by the saturation IRM value compared with that of the leucosomes (Figure 6-7b).

For melanosomes, the hysteresis loop suggests a much higher content of paramagnetic minerals (biotite and other iron-bearing minerals) and some ferromagnetic minerals. The thermomagnetic curve shows that the bearing ferromagnetic mineral could be magnetite, while the IRM acquisition curve indicates some high-coercivity minerals, such as goethite and hematite (Figure 6-7c).

More data can be found in the supplemental information Figures 6-S1 and 6-S2.

6.5 Anisotropy of Magnetic Susceptibility

Except for samples from Sites 126-129, which could be related to a later dioritic pluton, all studied samples show a consistent magnetic fabric with vertical poles of magnetic foliation (K3) and randomly distributed horizontal magnetic lineations (K1) after putting the rock layers to horizontal (Figure 6-8).

The average anisotropy degree of studied samples ($P_J = 1.198 > 1.05$, $n = 206$; Figure 6-8a) indicates a tectonic fabric. Meanwhile, the P_J value show a significant increase from leucosome ($P_J = 1.169$, $n = 63$; Figure 6-8b) to mesosomes ($P_J = 1.195$, $n = 71$; Figure 6-8c) to melanosomes ($P_J = 1.232$, $n = 72$; Figure 6-8c). In addition, only 1.4% melanosomes have prolate shape parameter (T), while in the mesosomes and leucosomes, the proportion of prolate samples increase to 4.2% and 11.1%, respectively.

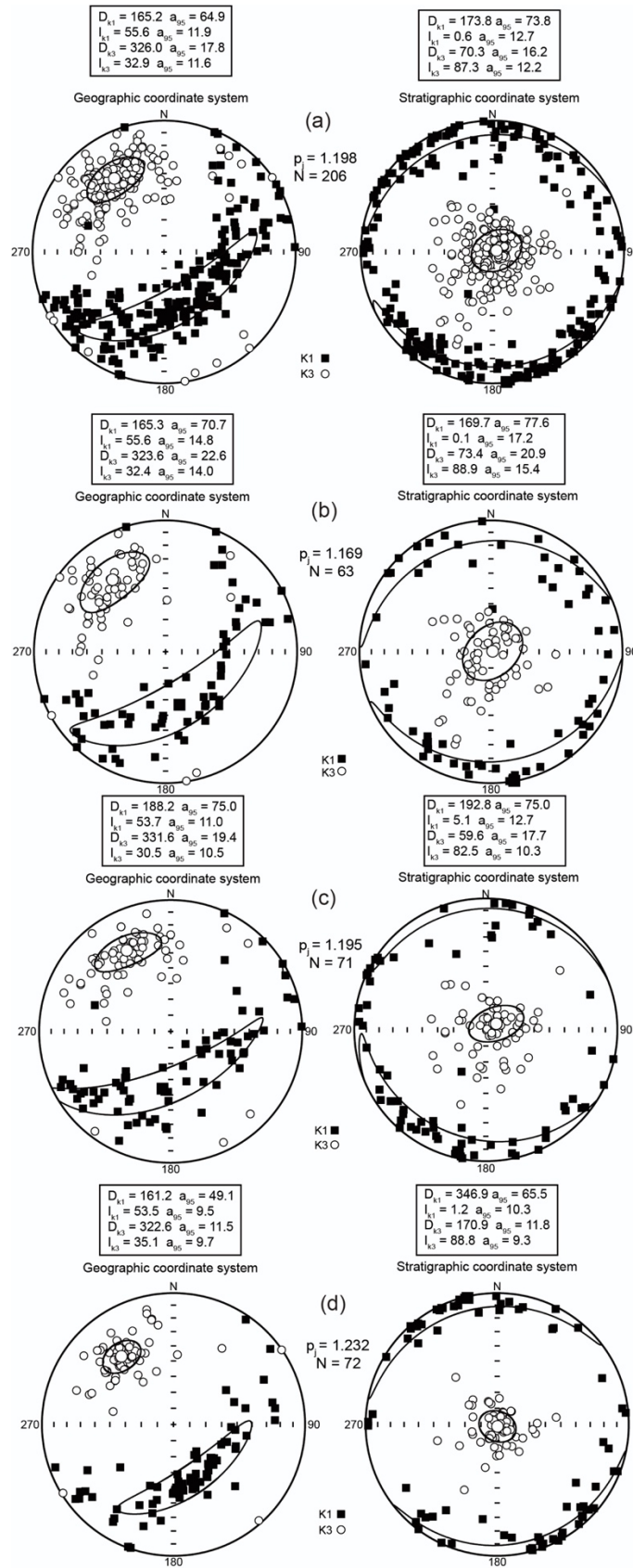


Figure 6-8. Anisotropy of magnetic susceptibility results of (a) all studied samples; (b) leucosomes; (c) Mesosomes; and (d) melanosomes.

6.6 Paleomagnetism

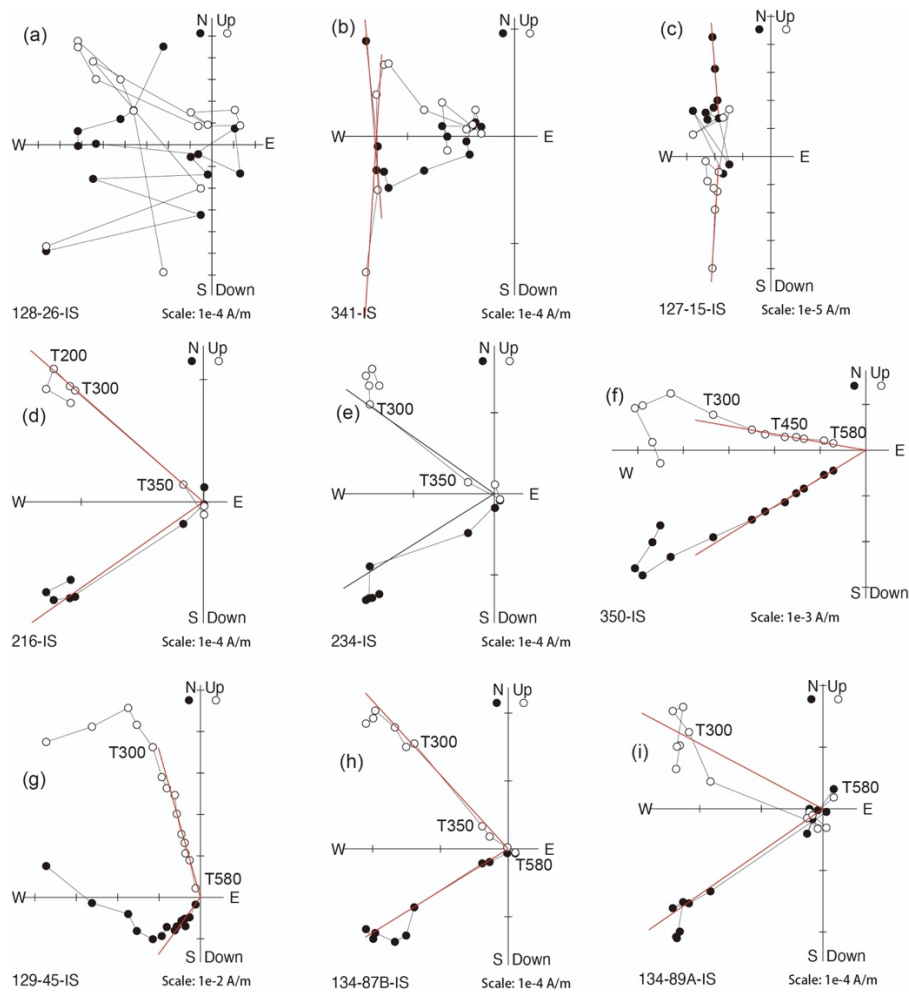


Figure 6-9. Representative demagnetization projections: (a) erratic result that is unable to analyze; (b-c) low-temperature components; (d-e) high-temperature components isolated by 350 °C; (f-i) high-temperature components isolated by 580 °C. Note: (G) specimens 129-45 is from meta-dioritic migmatite in section B.

Most paleomagnetic specimens show erratic demagnetization curves and are thus unable to isolate meaningful magnetic components (Figure 6-9a). However, around one-third of the specimens reveal significant components. Sixty-two low-temperature components were removed by ~200-250 °C (Figures 6-9b-c). Their fisher mean direction is $D_g = 348.3^\circ$, $I_g = 59.8^\circ$, $\alpha_{95S} = 4.9^\circ$, and $n = 62$ in the geographic coordinates ($_g$), being close to the present earth field (Figure 6-10a); and is $D_t =$

135.9°, $I_t = 72.0^\circ$, and $\alpha_{95t} = 9.2^\circ$ in the tectonic coordinates (t) (Figure 6-10b, red dashed oval), which is more dispersive. According to the field observation, magnetic properties, and the AMS results, samples from Sites 126-129 are collected from a later intruding dioritic pluton and thereby are unsuitable to perform tilt-correction using the foliations. Therefore, mean direction of only sediments-derived samples in the tectonic coordinates is calculated as: $D_t = 121.6^\circ$, $I_t = 51.8^\circ$, $\alpha_{95t} = 10.6^\circ$, and $n = 37$ (Figure 6-10b, blue oval).

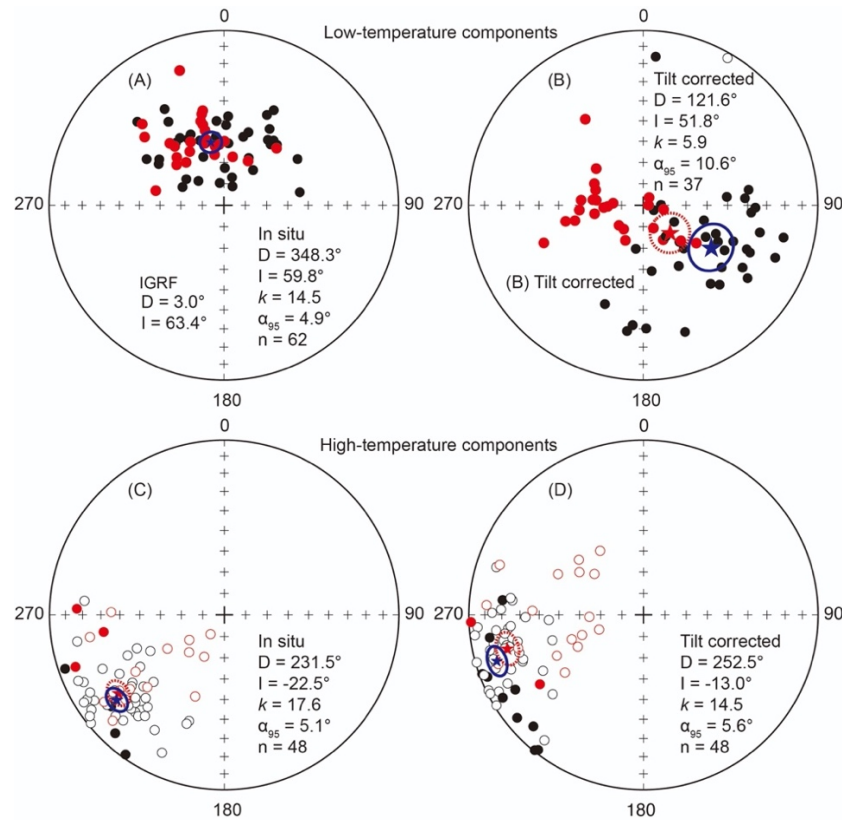


Figure 6-10. Fisher mean directions at specimen-level for (a-b) low-temperature components in the geographic and tectonic coordinates, respectively; and for (c-d) high-temperature components in the geographic and tectonic coordinates, respectively. Note: Red spots (hollow and full) mean ChRMs for specimens from Sites 126-129; Red dashed ovals stands for the mean directions for all specimens (attention: in Figure 6-10a, blue oval represents all specimens' Fisher mean direction), while blue ovals refer to mean directions for all specimens excluding that from Sites 126-129.

Some specimens' magnetic remanence tends to linearly decay to the origin by ~350 °C after removing low-temperature component (Figures 6-9d-e), while most specimens decay to the origin at ~580 °C (Figures 6-9f-i). Totally, 67 ChRM directions were isolated and their fisher mean directions are $D_g = 233.5^\circ$, $I_g = -25.6^\circ$, $\alpha_{95s} = 5.5^\circ$, and $n = 67$ (Figure 6-10c, red dashed oval), and $D_t = 256.0^\circ$, $I_t = -20.7^\circ$, and $\alpha_{95t} = 6.4^\circ$ (Figure 6-10d, red dashed oval). Furthermore, the fisher mean directions without specimens from Sites 126-129 are $D_g = 231.5^\circ$, $I_g = -22.5^\circ$, $\alpha_{95s} = 5.1^\circ$, and $n = 48$ (Figure 6-10c, blue oval), and $D_t = 252.5^\circ$, $I_t = -13.0^\circ$, and $\alpha_{95t} = 5.6^\circ$ (Figure 6-10d, blue oval).

6.7 Preliminary Discussions

6.7.1 Evolution history of the Baluntai Migmatites

The zircons from the Baluntai migmatitic granite show a typical core-rim structure. According to the U-Pb zircon dating result of the cores, the protolith of Baluntai migmatitic granite has a Precambrian age clustering at 917 ± 11 Ma (Figure 6-5a). In addition, the detrital zircons selected from the meta-greywacke also have ages ranging from Neoproterozoic to Mesoarchean (Figures 6-5b-c). These data indicate that the Chinese Central Tianshan is a Precambrian microcontinent. Previous studies suggest that the Chinese Central Tianshan have a tectonic affinity with the Tarim Block (Zhai et al., 2015; Wang et al., 2018).

The zircon rims of the migmatitic granite sample have a Late Devonian (Frasnian) metamorphic age at 377.3 ± 5.1 Ma. This Frasnian metamorphism could be related to the emplacement of the migmatitic granite. Combining with regional geology, the amalgamation/collision between Central Tianshan and South Tianshan due to the closure of major back-arc basin(s) could result in this emplacement, i.e., because of the crustal thickening, the Frasnian anatexis occurred in the Central Tianshan area.

The monazite dating results reveal another high-temperature metamorphic event at

~310 Ma. Combining with the ~377 Ma event discussed above, we could propose two tectonic evolution scenarios for the Central Tianshan: 1. Continuous high-temperature metamorphism from at least the Frasnian (~377 Ma) to Moscovian (~310 Ma), as a result of the diachronous amalgamation of back-arc basins in the South Tianshan region (Wang et al., 2018); 2. After the Frasnian event, another high-temperature metamorphism happened in the Moscovian, probably triggered by large-scale lateral displacements in the SW CAO (since the initiation times of the Nalati Fault, North Tianshan Fault - Middle Tianshan Shear Zone, and Baluntai Fault are all around 310 Ma; see geological background section; He et al., 2020).

The muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dating result of the migmatitic granite gives a muscovite cooling age (blocking temperature: ~300-350 °C) as ~310 Ma, and the biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dating result of the melanosomes also shows a biotite cooling age (blocking temperature: ~250-300 °C) at ~310 Ma. Associated with the ~310 Ma monazite U-Th-Pb age, considering the confidence interval, we propose a fast exhumation process for the Baluntai migmatite at the Moscovian. Since the fault system in this region was initiated at ~310 Ma, we propose that the first tectonic scenario is more plausible.

6.7.2 “Primary” magnetic fabric and secondary magnetic remanence

Except for samples from Sites 126-129, all studied samples show vertical poles of magnetic foliation (K3) and randomly distributed horizontal magnetic lineations (K1) after foliation correction (Figure 6-8).

The average anisotropy degree of studied samples ($P_J = 1.198 > 1.05$, $n = 206$; Figure 6-8a) indicates a tectonic fabric. The P_J value shows a significant increase from leucosomes ($P_J = 1.169$, $n = 63$; Figure 6-8b) to mesosomes ($P_J = 1.195$, $n = 71$; Figure 6-9c) to melanosomes ($P_J = 1.232$, $n = 72$; Figure 6-8d). Moreover, only 1.4% melanosomes have prolate shape parameter (T), while in the mesosomes and leucosomes, the proportion of prolate samples increase to 4.2% and 11.1%, respectively.

The above AMS results suggest that these fabrics were acquired under vertical

principal stress (lithostatic pressure). The melanosomes and mesosomes are more deformed as suggested by higher P_J , T , and microscopic observation (Figures 6-4c-d); while the leucosomes are less deformed with lower P_J values, and exhibit a magmatic structure under a microscope (Figures 6-4a-b). The relatively concentrated lineation in melanosomes may have inherited from the protoliths and/or paleosome. Therefore, we regard the AMS was acquired during the latest migmatization process (~310 Ma) with a horizontal attitude, and thus be thought as “primary” AMS for the Baluntai migmatite. Horizontal fabrics are usually formed during the transition from orogenic thickening of crust to the post-orogenic gravitational collapse, which is a relatively stable tectonic stage without compression/extension stress. With this, we suggest such a tectonic transition setting for the Baluntai migmatites and that the convergence and amalgamation of the Chinese Central Tianshan and South Tianshan terminated by Moscovian.

The Fisher mean direction without specimens from metadiorite-derived Sites 126-129 (since the paleohorizontal plane is unable to know for metadiorite-derived migmatites) are $D_g = 231.5^\circ$, $I_g = -22.5^\circ$, $\alpha_{95s} = 5.1^\circ$, and $n = 48$ (Figure 6-10c, blue oval), and $D_t = 252.5^\circ$, $I_t = -13.0^\circ$, and $\alpha_{95t} = 5.6^\circ$ (Figure 6-10d, blue oval).

The ~310 Ma monazite U-Th-Pb and Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dating results suggest that the age of magnetic remanence should also be ~310 Ma. Based on the fast exhumation process, one may put a question that if our paleomagnetic data can fully represent the paleogeomagnetic secular variation. Even though no fold test is available, the statistical analysis using Tauxe et al. (2004) and Deenen et al. (2011) criteria and dominantly reverse polarity of the ChRMs suggest a full representation of the paleogeomagnetic secular variation.

However, we found that our paleomagnetic result is inconsistent with previously published regional latest Carboniferous-Early Permian paleomagnetic data. For example, the highest paleolatitude we can obtain using our paleomagnetic data is around 14.6°N (in the geographic coordinates). In contrast, the average paleolatitude

for blocks in the SW CAO (Tarim, Yili, South Junggar, Turpan-Hami) is around 34° N (Zhu et al., 2020). However, lines of geological evidence indicate that the Central Tianshan probably had a similar paleolatitude as neighboring blocks at the Moscovian.

We suppose that the magnetic remanence was obtained after a tilting of the foliations, probably with a NW-SE strike (present coordinates, parallel with the tectonic lines in the region), and the Baluntai migmatites at least underwent a counterclockwise rotation along a horizontal SE rotation axis up to ~30° since the Moscovian. P-T condition of 440-460 Ma granodiorites was estimated by Zhong et al. (2019, thesis), suggesting that the north pluton had greater depth. It is to say that more significant uplift and exhumation occurred in the north, which is consistent with our proposition. Moreover, the latest Carboniferous and earliest Permian granitic plutons intruded to the north and south of the studied migmatites, contributing to the tilting of the migmatites (personal communication with Bo Wang).

The “primary” AMS and secondary remanence carried by the Baluntai migmatites could be a result of the low content of ferromagnetic minerals (*sensu lato*). To modify the AMS, the geological events should fulfill three conditions: 1. temperature; 2. stress; 3. sufficient duration. In our case, the temperature and duration would not be enough to modify primary AMS which is mainly carried by paramagnetic minerals, chiefly biotite. However, to change magnetic remanence carried by the magnetite, it is easier since it only needs to reach a temperature where the magnetic domains reorient in the magnetite (perturb magnetic domain Pseudo-Single-Domain & Multi-Domain magnetite) instead of reorienting its crystalline prefer orientation (Single-Domain magnetite). Furthermore, since the content of ferromagnetic minerals is relatively low (Figure 6-S3), even if some ferromagnetic minerals (SD magnetite) reorient their crystalline prefer orientation, it is overwhelmed by the magnetic susceptibility of abundant biotite.

6.8 Summary

- (1) Under a binocular microscope, a typical magmatic structure was found in the

leucosomes and mesosomes; however, a moderate degree of deformation was observed in the melanosomes.

(2) The crystallization age of the migmatitic granite was 377.3 ± 5.1 Ma according to zircon U-Pb dating; the meta-greywacke sample has a youngest U-Pb detrital zircon age peak at ~ 312 Ma. Monazite selected from melanosomes gave U-Th-Pb ages at 309–312 Ma. Moreover, $^{40}\text{Ar}/^{39}\text{Ar}$ dating performed on muscovite and biotite selected from migmatitic granite and melanosome showed plateau ages over all the steps at 310.5 ± 6.5 Ma and 310.8 ± 7.6 Ma, respectively.

(3) The migmatization in the Baluntai area occurred at ~ 377 Ma and probably continues to ~ 310 Ma, which likely resulted from diachronous closure of the back-arc basin(s) between the Central Tianshan and the South Tianshan. At ~ 310 Ma, the Baluntai migmatites underwent a fast uplift/exhumation.

(4) Based on the AMS results, the migmatites preserve a “primary” horizontal magnetic fabric acquired during the migmatization, indicating a relatively stable tectonic stage related to the transition from orogenic thickening of crust to the post-orogenic gravitational collapse. Hereby, we suggest the convergence and amalgamation of the Central Tianshan and South Tianshan terminated by Moscovian.

(5) In contrast, the ChRM isolated from the migmatites is secondary, whose age is probably at ~ 310 Ma. We attribute this phenomenon to different acquisition mechanisms between AMS and magnetic remanence, and the low content of ferromagnetic minerals in studied rocks.

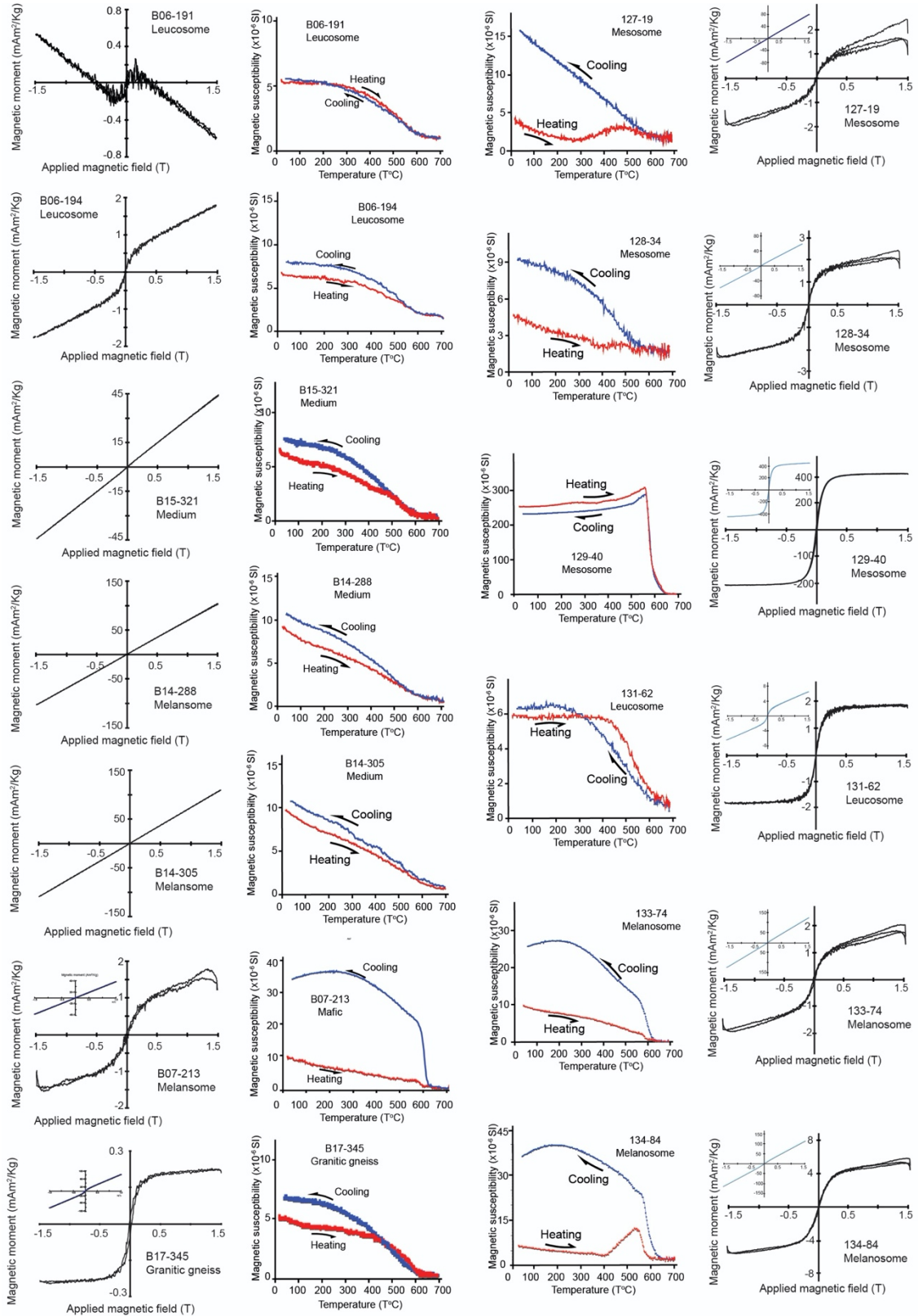


Figure 6-S1. Supplemental hysteresis loops and thermomagnetic curves of the leucosomes, mesosomes (medium), and melanosomes (mafic) from the Baluntai migmatites.

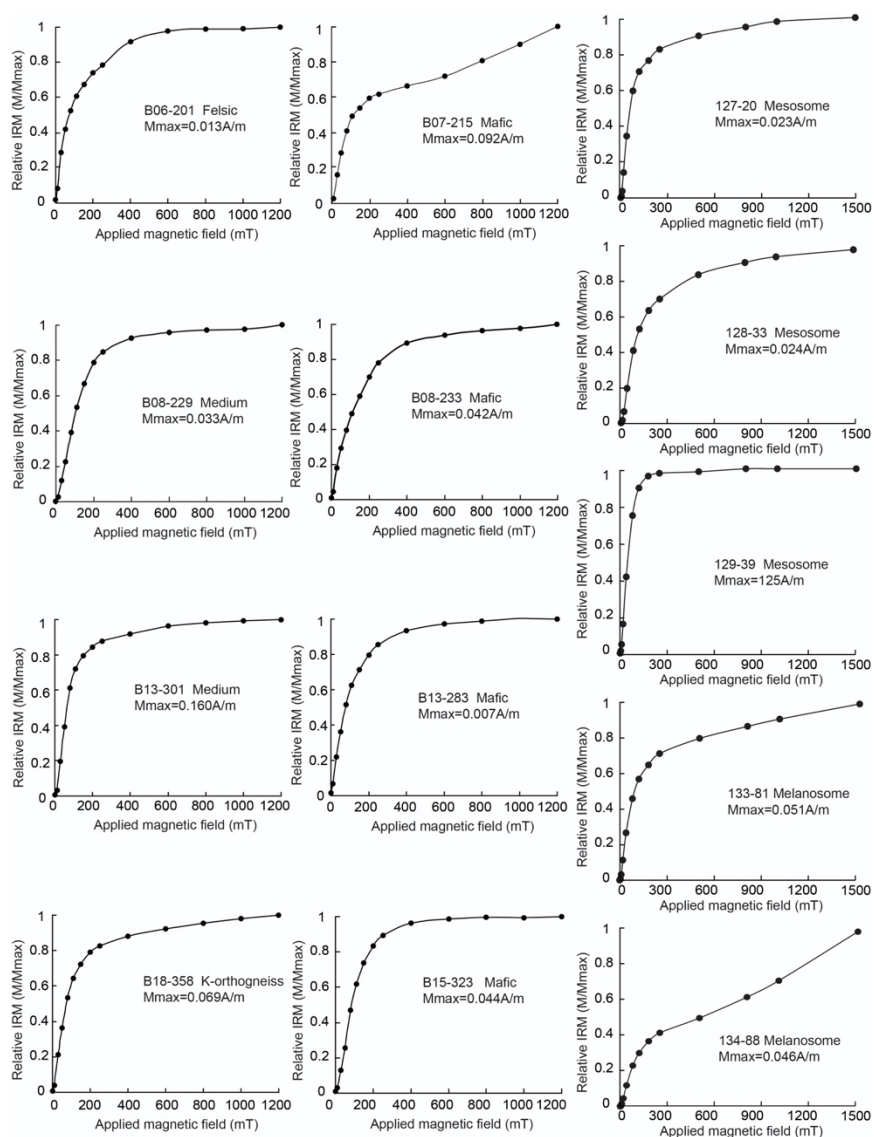


Figure 6-S2. Supplemental Isothermal Remanent Magnetization acquisition curves of the leucosomes, mesosomes (medium), and melanosomes (mafic) from the Baluntai migmatites.

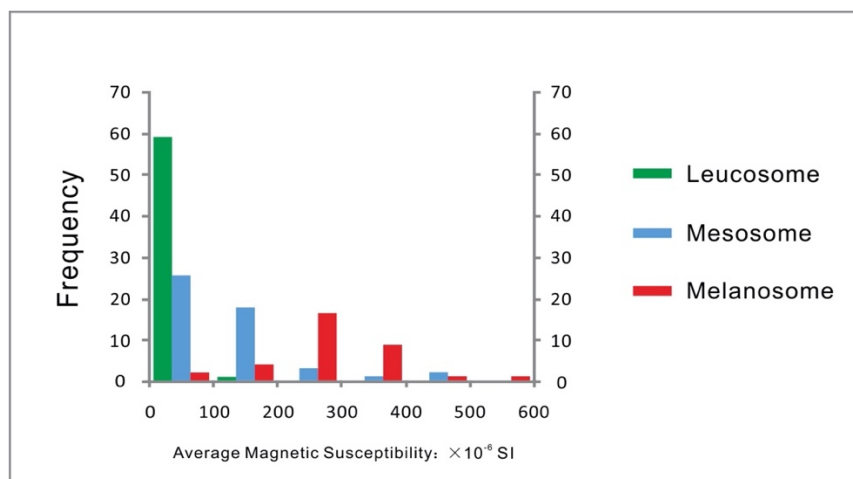


Figure 6-S3. The distribution of magnetic susceptibility of the Baluntai migmatites.

Chapter 7. Setting up the Early Permian Beishan Paleomagnetic ‘Bridge’ for the southern Central Asian Orogenic Belt

In the Beishan Belt, different latest closure times of the Paleo-Asian Ocean have been suggested (see section 1.2). The limited previous research approaches, which are mainly geochronology and geochemistry of igneous rocks and sometimes lead to ambiguous tectonic interpretations, could be a main origin of such a controversy. Therefore, more constraints from different perspectives are essentially required (e.g., Chen et al., 2016; Niu et al., 2018).

Paleomagnetism plays a key role in resolving such puzzles. According to previous paleomagnetic studies, Zhao et al. (2020) proposed that the Mongolia Block should have been amalgamated with the North China Block at least in the P₁, while Zhang et al. (2018) and Ren et al. (2020) indicated that the PAO closed at the end of Permian. No significant or weak latitudinal relative motion has taken place among the major blocks in the Chinese Tianshan areas since the late Carboniferous (Wang et al., 2007; Zhu et al., 2018, 2019). Furthermore, Xu et al. (2019) suggested that the PAO was still open in the P₁. Critically located between the SW and SE CAOB (Figure 7-1a), the Beishan Belt provides such an opportunity for testing the late Paleozoic-early Mesozoic tectonic transition process of the CAOB, but lacks appropriate paleomagnetic data so far. In view of this, we carried out a paleomagnetic study on the Gubaoquan doleritic dike swarm in the Beishan Belt along the southern boundary of the CAOB.

7.1 Geological background and sampling

The Shuangyingshan terrane consists of possible Precambrian basement, Paleozoic clastic rocks and carbonate, mafic-ultramafic and granitic intrusions, and volcanic rocks. Granitic rocks with different geochemical characteristics, including arc-related and post-collision extensional, intruded between 437 to 217 Ma (He et al., 2018b). The

mafic-ultramafic rocks were generally emplaced during 295 to 260 Ma in an extensional setting (Su et al., 2010; Zhang et al., 2011). The volcanic rocks formed in an arc setting during 451-367 Ma (Guo et al., 2014). In the southernmost part of this terrane, medium- to high-grade metamorphic rocks, mainly including gneisses, enclosing retrogressed Ordovician eclogite boudins (Soldner et al., 2020), quartz schists, and migmatites, crop out in the Gubaoquan area along a 10-km-wide belt. To the south, the P₁ strata of the Liuyuan unit consist of generally fining-upwards sedimentary sequences and upper thick packages of basalts (Chen et al., 2016).

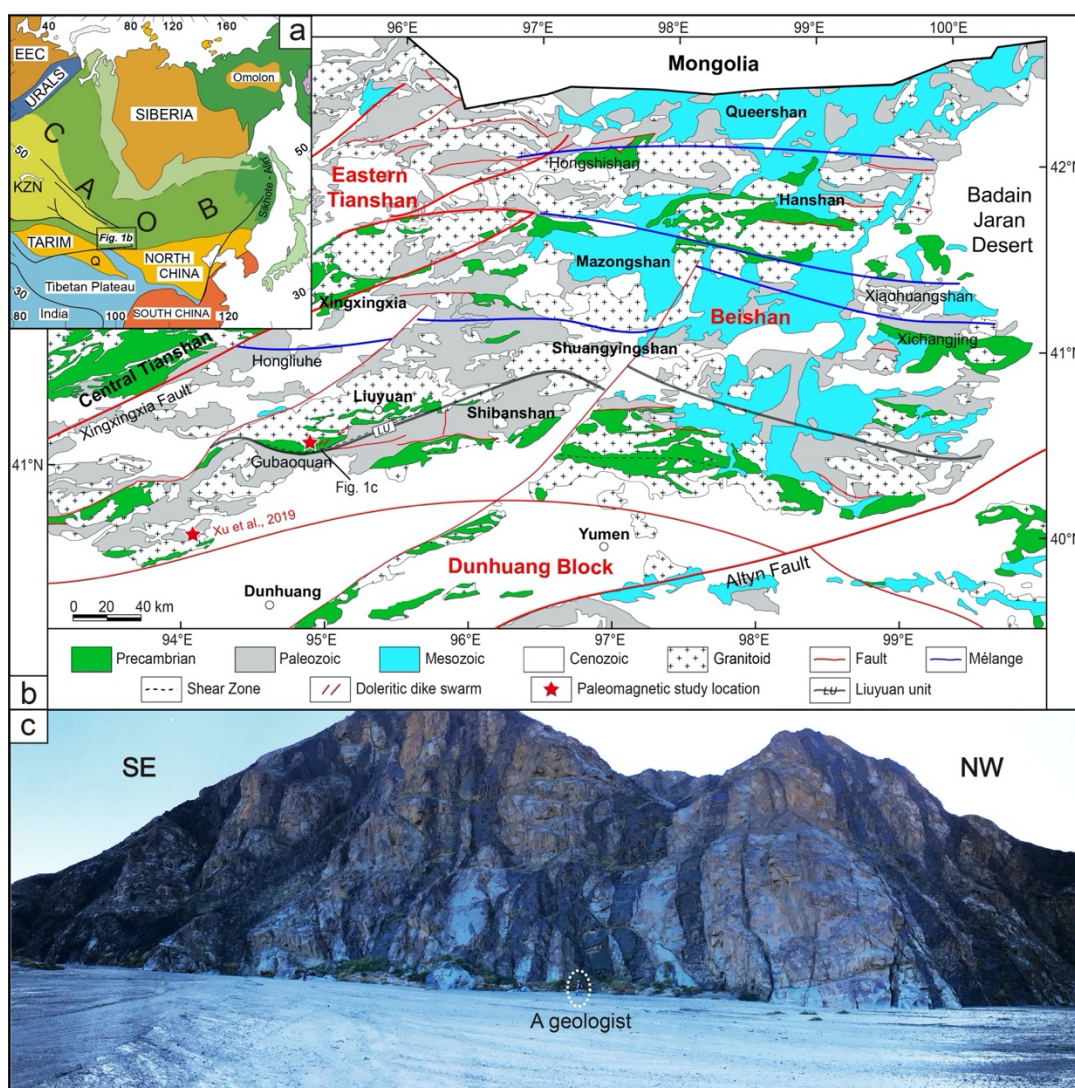


Figure 7-1. (a) Simplified tectonic division of the eastern Asia (after Jahn, 2004). EEC = East European Craton; KZN = Kazakhstan; Q = Qaidam. (b) Geologic map of the Beishan Belt (after He et al., 2018b). (c) A field photo of the studied doleritic dike swarm in the Gubaoquan area.

Doleritic dike swarms are widespread in the Gubaoquan area, intruding in the Precambrian gneisses and Silurian-Devonian foliated granites with NEN-SWS to ENE-WSW (40° – 60°) strikes and nearly vertical dips; the dike emplacement age was constrained as 282 ± 6 Ma based on two youngest zircons (Zhang et al., 2015). Our sampling section is in this area, where numerous subvertical fresh doleritic dikes striking NEN-SWS intrude in Neoproterozoic augen gneisses (Yuan et al., 2015; Figures 7-1c and 7-2). Sixty-five paleomagnetic cores were drilled from 20 dikes with widths of 1-10 m (1 to 10 cores per dike according to its width), and 11 cores were taken from country rocks for the baked contact test (Table 7-1). All cores were oriented by both magnetic and solar compasses, whenever possible.

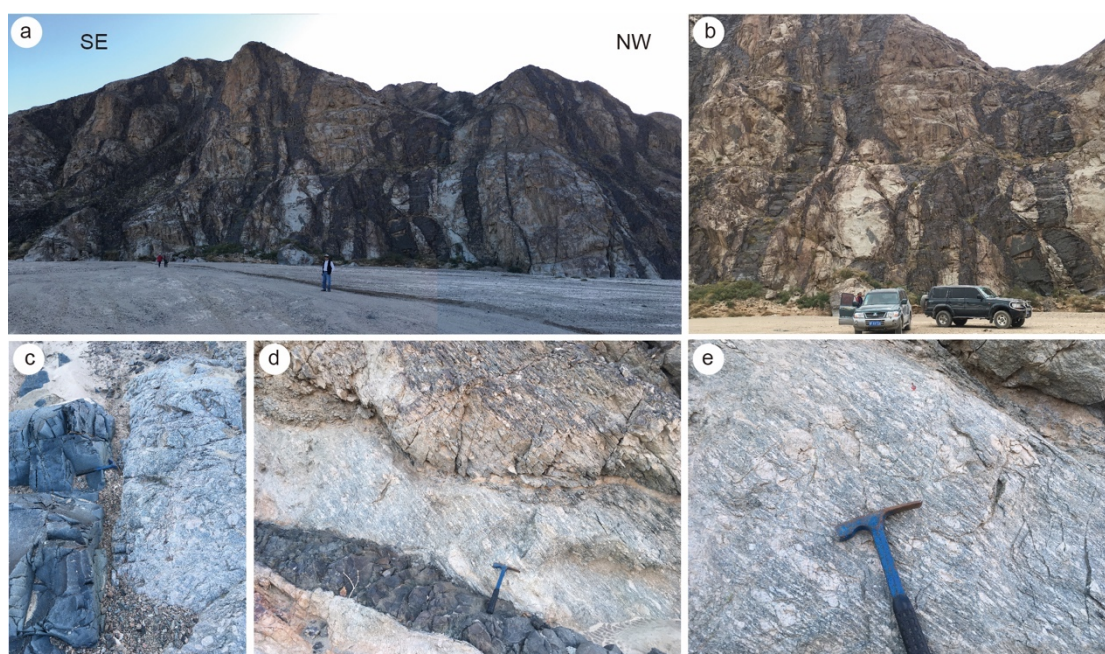


Figure 7-2. Field photos of the sampling section in the Gubaoquan area, south Beishan: (a and b) a general view of the doleritic dikes intruding into the Neoproterozoic orthogneisses; (c and d) sharp contact relationship between the dikes and orthogneisses; (e) augen structure of the orthogneisses.

7.2 Rock magnetism and AMS analyses

Thermomagnetic experiments show obvious increases of magnetic susceptibility during heating at 20-280 °C and sometimes at 400-560 °C, significant decreases at 280-400 °C and sharp ones at 560-600 °C, and occasionally tiny decreases at 600-700 °C; much higher magnetic susceptibility is often observed on cooling curves than heating ones (Figures 7-3a-c and 7-4). Room-temperature hysteresis loops for most doleritic samples show “pot-bellied” features (Huang et al., 2017) with relatively low coercivity and remanence coercivity (Figures 7-3d, 7-3f and 7-4). Loops of other doleritic and all orthogneissic samples indicate soft-coercivity magnetic mineral(s) (Figures 7-3e, 7-3g and 7-4).

Therefore, magnetite seems the principal magnetic mineral with some pyrite, pyrrhotite and/or hematite. Possible multi-phase mineral transitions happened during heating, including the ones from pyrite to pyrrhotite and magnetite. After plotting hysteresis parameters of 17 samples on the Day plot (Dunlop, 2002), they drop in the region with mixtures of single-domain (SD) and multi-domain (MD) magnetites and close to the SD region (Figure 7-3h).

AMS results of doleritic specimens present a low anisotropy degree ($P_J < 1.05$; Figure 7-3i). The poles of the magnetic foliation are nearly vertical, while magnetic lineations are divided into two groups: 86% are striking NEN-SWS, parallel to the dike strike, and 14% are WNW-ESE (Figure 7-3j). The magnetic fabrics of the country rocks (orthogneissic specimens) are scattered (Figure 7-3k).

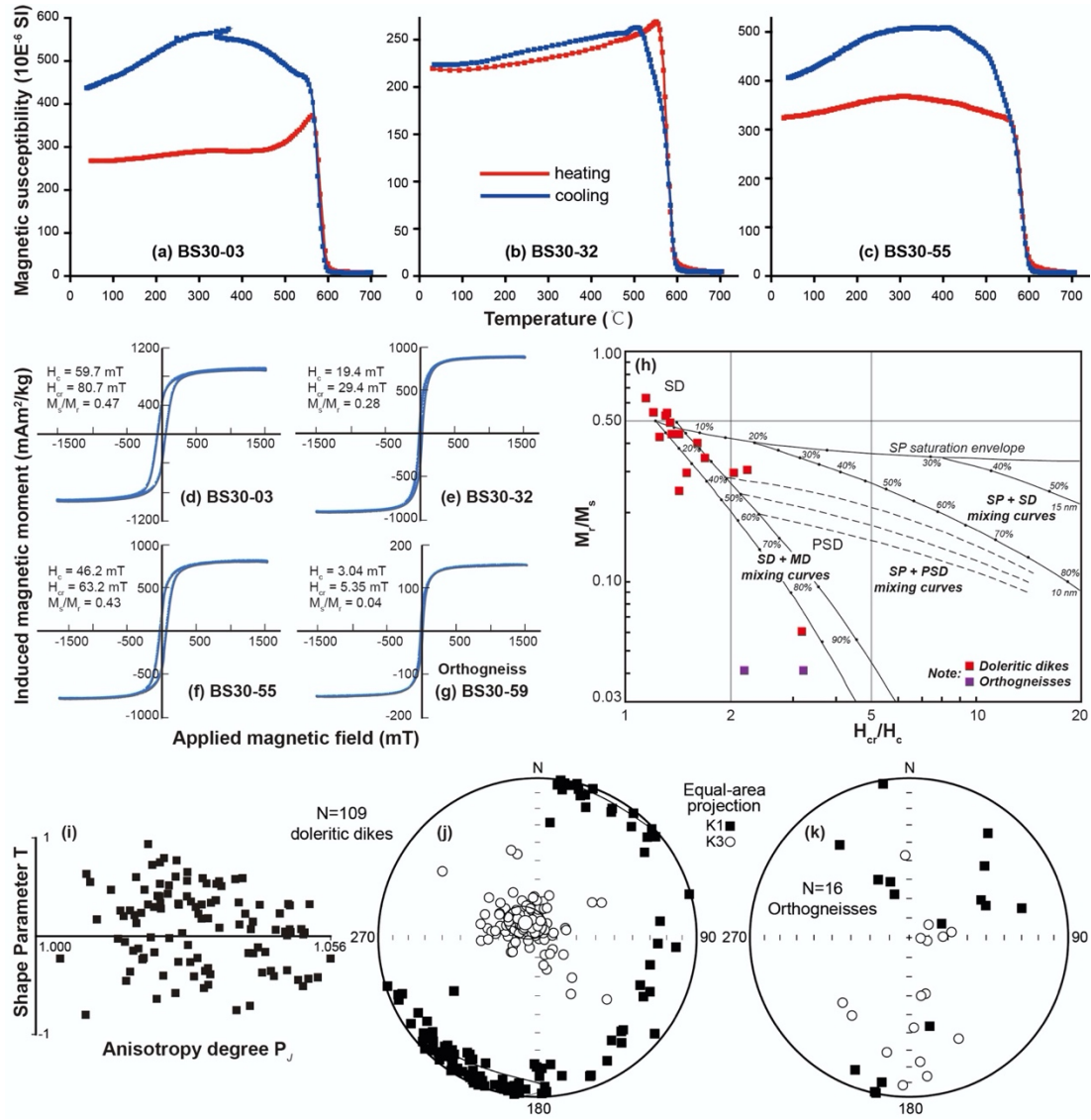


Figure 7-3. (a-c) Thermomagnetic curves and (d-g) hysteresis loops of representative samples. (h) Day plot of 17 representative samples. (i-j and k) AMS results of doleritic dikes and orthogneisses, respectively.

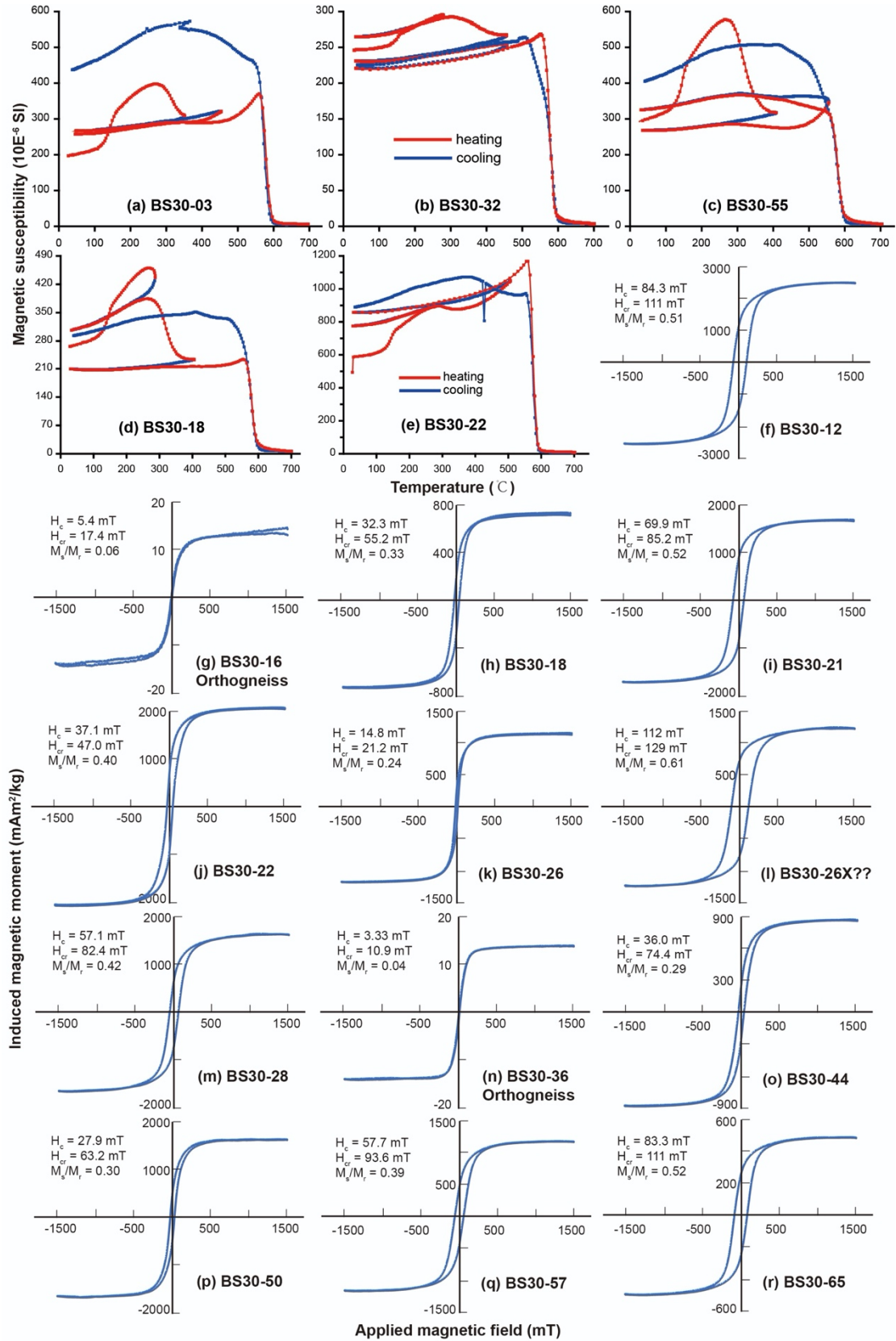


Figure 7-4. Supporting thermomagnetic curves and magnetic hysteresis loops.

7.3 Scanning electronic microscope (SEM) observation

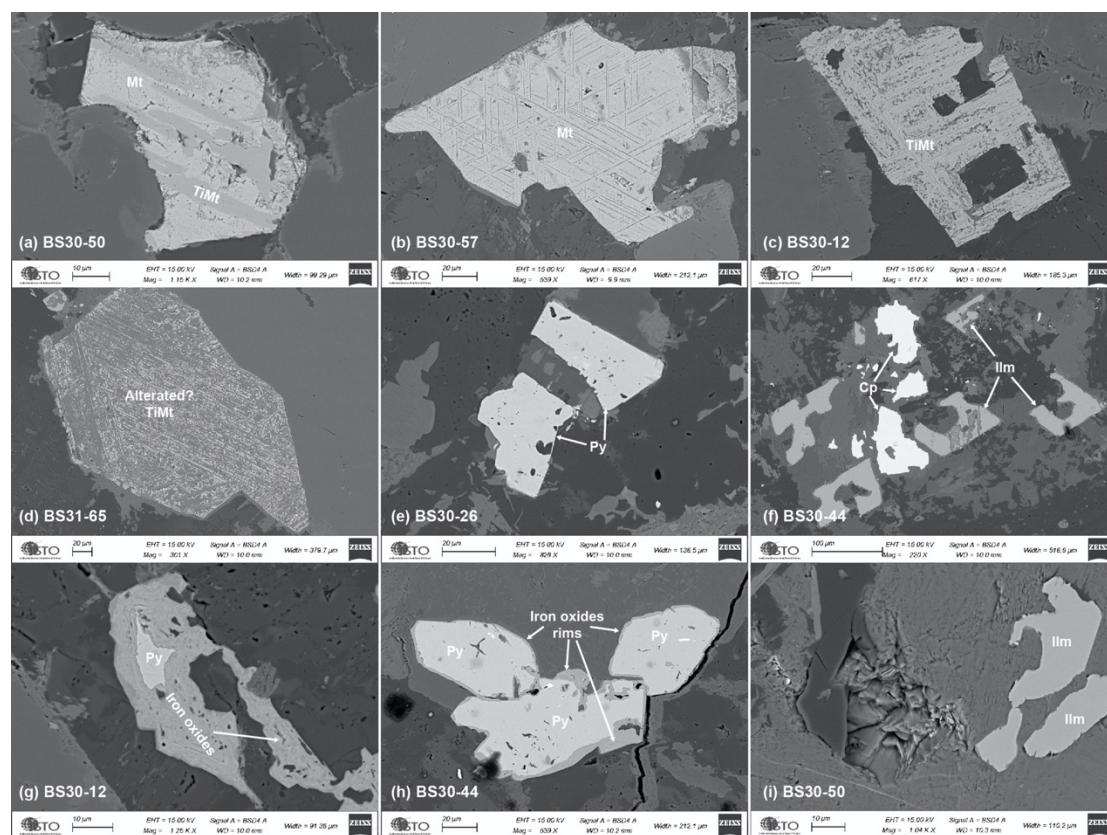


Figure 7-5. Scanning electronic microscope observations: (a-b) irregular magnetite (Mt) and titanomagnetite (TiMt); (c-d) possibly altered titanomagnetite; (e-f) irregular pyrite (Py), chalcopyrite (Cp), and ilmenite (Ilm); (g-h) iron oxides rims of pyrite; and (h-i) cracks and pores without authigenic minerals.

Irregular MD magnetites and titanomagnetites (30-150 μm) usually do not show alteration (Figures 7-5a-b), only minor titanomagnetites with diameters of 50-200 μm experienced exsolution and possibly strong alteration (also might be due to excessive polishing; Figures 7-5c-d). Abundant irregular pyrite, and minor ilmenite and chalcopyrite are found (Figures 7-5e-f). Pyrite occasionally shows iron oxides rims (Figures 7-5g-h). No authigenic mineral is discovered within fractures and pores (Figures 7-5h-i). Furthermore, typical pyrrhotite or hematite is not observed, possibly due to their low contents and/or tiny sizes. It is probably because of the mixture of

magnetite/titanomagnetite with some high-coercivity minerals (e.g. iron oxides rims of pyrite) that our samples plot close to the SD region in the Day plot (Figure 7-3h). In this case, MD/Pseudo-single-domain magnetites are the principal magnetic mineral. Corresponding analytical results from the energy-dispersive X-ray spectrometer are shown in Figure 7-6.

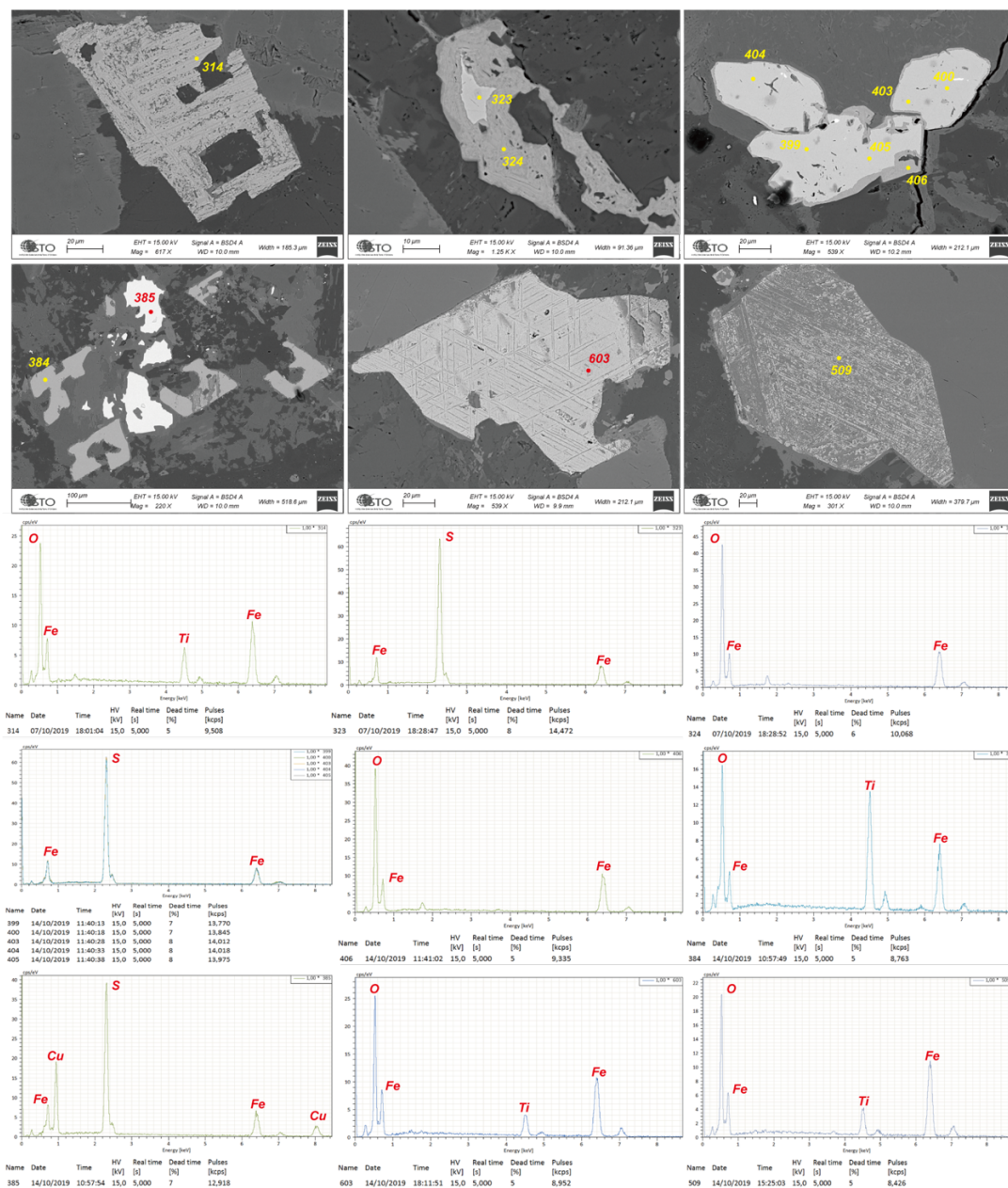


Figure 7-6. The energy-dispersive X-ray spectrometer results corresponding to the SEM observations.

7.4 Plagioclase $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology

Three $^{40}\text{Ar}/^{39}\text{Ar}$ dating runs were performed on plagioclase selected from Sample BS30-32 on the margin of a 10 m-width doleritic dike (Figures 7-7; Table 7-2). Run N216 shows a quasi-plateau age of 283.8 ± 1.8 Ma over the last steps (70% ^{39}Ar released) after some irregular initial steps (Figure 7-7). In contrast, runs N218 and N219 exhibit initial concordant ages of 259.9 ± 2.6 and 265.6 ± 3.0 Ma respectively at low temperature (79% and 67% ^{39}Ar released), followed by a common older age over the last steps (16% and 33% ^{39}Ar released, Figure 7-7).

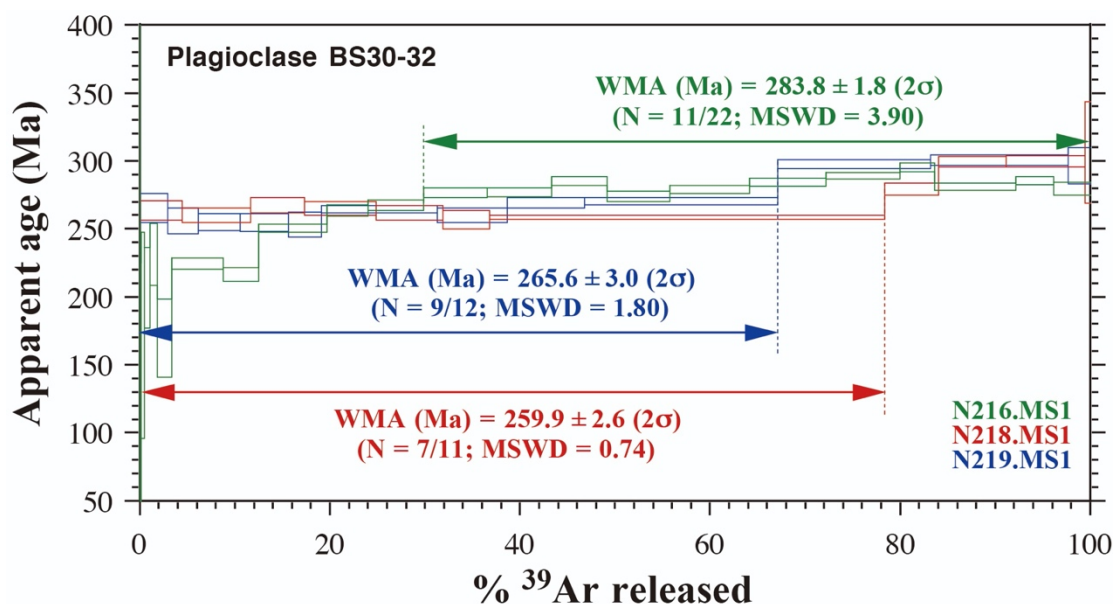


Figure 7-7. $^{40}\text{Ar}/^{39}\text{Ar}$ geochronologic results of plagioclase of Sample BS30-32.

7.5 Paleomagnetism

Low-coercivity/temperature components of forty-one samples were removed by ~ 10 -25 mT/ ~ 200 -250 $^{\circ}\text{C}$, and their directions are close to the present earth field (Figure 7-8).

Most AF demagnetized specimens show a high-coercivity (up to 45-70 mT) Characteristic Remanent Magnetization (ChRM) component trending directly to the origin (Figures 7-9a-b). However, a few AF demagnetized specimens show a stable remanent component which trends to the origin at middle-field steps, but deviates

from the origin at high-field steps (normally >60 mT; Figure 7-9c). We consider the component at middle-field steps as ChRM and the high-field component as a mixture of ChRM and secondary remanence carried by high-coercivity minerals like iron oxides rim of pyrite (Figures 7-5g-h). Concerning the thermal demagnetized specimens, their magnetic remanence tends to linearly decay to the origin after removing low-temperature component (Figure 7-9d), but the remanent direction of most specimens behaves erratically at high-temperature steps (Figures 7-9e-f), which probably results from abundant newly-formed magnetite during heating (Figures 7-3a and 3c). Therefore, careful comparisons between AF and thermal demagnetizations were made to get representative ChRM directions (Figures 7-9b and 9e). Consequently, ChRM direction for each dike were determined and a dike-mean direction was obtained: $D_g = 174.9^\circ$, $I_g = -51.4^\circ$, $\alpha_{95} = 4.8^\circ$, and $n = 20$ dikes (Figure 7-9g; Table 7-3).

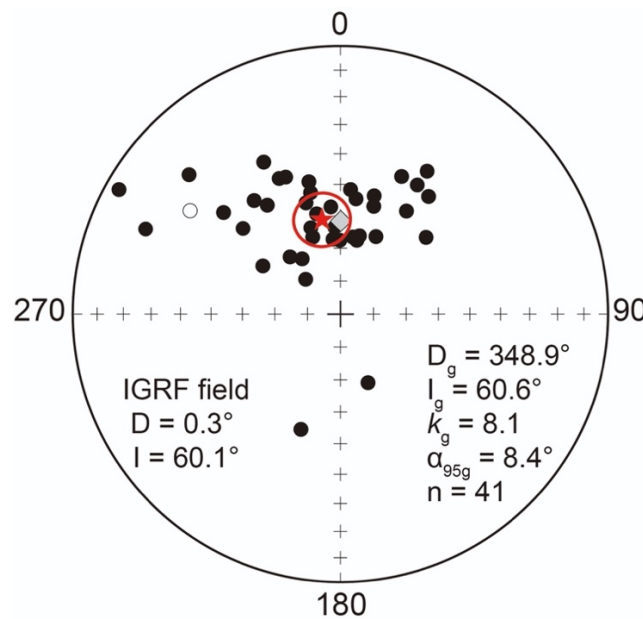


Figure 7-8. Magnetic directions of low-temperature/ low-coercivity components isolated from demagnetized samples and the comparison between their mean-direction (star) and present earth field (diamond).

Furthermore, ChRMs were also isolated from orthogneissic specimens collected from the country rocks close to the dikes, and show significantly different directions from that of the neighboring doleritic specimens (e.g. Figure 7-9h; Table 7-3).

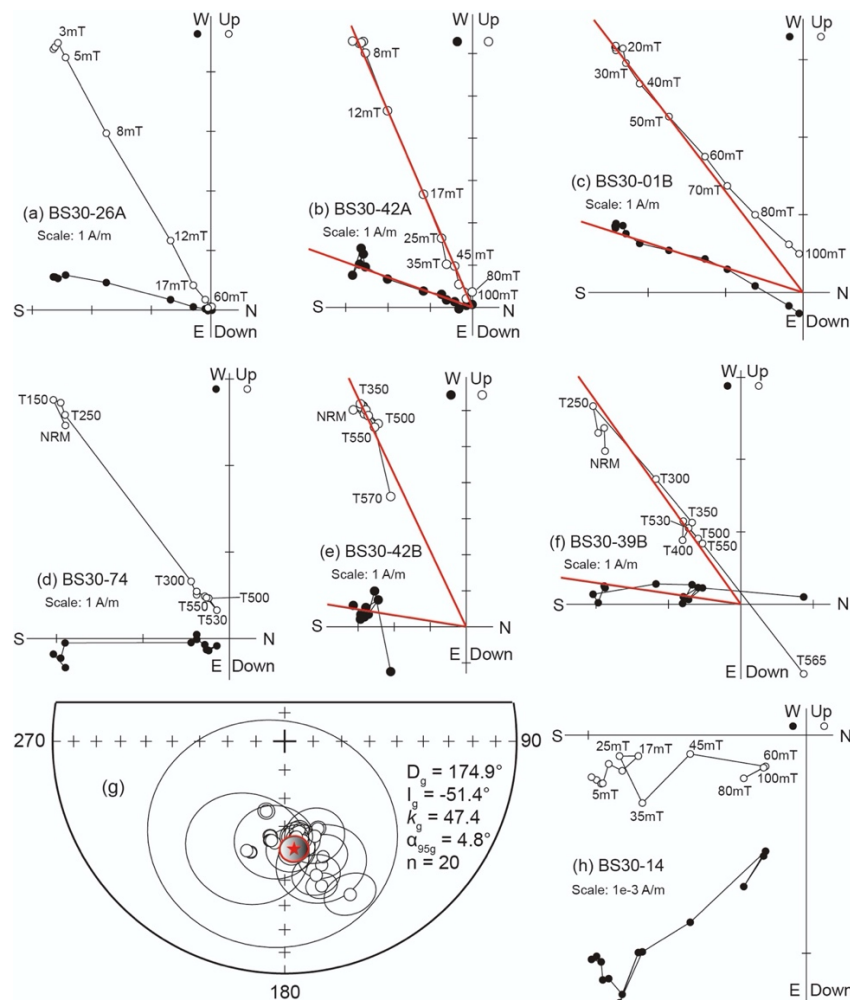


Figure 7-9. Representative demagnetization curves in the geographic coordinates for specimens from (a-f) the doleritic dikes and (h) the orthogneisses; (g) dike-mean ChRM directions.

7.6 Discussion on acquiring an Early Permian pole for the Beishan Belt

The $^{40}\text{Ar}/^{39}\text{Ar}$ run N216 provided a 284 Ma age consistent with the U/Pb intrusion age of the dikes. The other runs gave younger plateau ages at ~260-266 Ma, and may reflect either later magma-thermal activity with local ^{40}Ar remobilization since coeval magmatism has been reported in south Beishan (e.g. in the Xuanwoling and Xijianquanzi areas; Su et al., 2010; Zhang et al., 2010), or differential cooling of

the plagioclases (due to sub-grain domainal behaviour). In either case, a P_1 age can be assigned to the timing t when the dikes cooled through the blocking temperature of magnetite and Ti-poor titanomagnetite ($>500\text{ }^{\circ}\text{C}$, Figures 7-3a-c). Furthermore, based on the weak/lacking interaction between dikes and country rocks indicated by the AMS pattern and magnetic remanent directions, and the sharp contact between them (Figure 7-2), we suggest that the dikes might have experienced a fast cooling and thus t could be $\sim 280\text{ Ma}$.

The isolated ChRMs from the doleritic specimens are probably primary based on the following arguments: abundant irregular MD magnetite/titanomagnetite without intense alteration (Figures 7-5a-b); no authigenic mineral within fractures and pores (Figures 7-5h-i; Jiao et al., 2019; Zhang et al., 2019); the “pot-bellied” hysteresis loops and SD-MD mixing positions of doleritic specimens’ hysteresis parameters in the Day plot (Huang et al., 2017; Figure 7-3h); the low P_J value of the doleritic dikes suggesting no intense posterior deformation (Figure 7-3i); a sole reverse polarity consistent with the Kiaman Reversed Superchron (Figure 7-9g); a positive baked contact test by ChRMs of country rocks (Table 7-3); a possible weak influence of posterior magma-thermal events, which only partially reset the $^{40}\text{Ar}/^{39}\text{Ar}$ system of plagioclase ($225\text{--}300\text{ }^{\circ}\text{C}$; Cassata et al., 2009) at $266\text{--}260\text{ Ma}$, was unable to blur the P_1 ChRMs carried by magnetite; and, notably, paleosecular variation is fully represented in our data according to the Deenen et al. (2011) criteria ($A_{95} = 5.4^{\circ}$ and $A_{95\text{min}}/A_{95\text{max}} = 3.6^{\circ}/12.4^{\circ}$ for $N = 20$; Figures 7-10a-b).

Although no overlying strata are preserved above the sampled dikes, we propose that the dikes have not experienced significant tilting, because generally their hosting Neoproterozoic orthogneissic basement is hard to be tilted, and the AMS results support a non/weak-tilting ($<15^{\circ}$) scenario. Accordingly, a P_1 paleomagnetic pole for the Beishan Belt is calculated at $\lambda/\varphi = 80.9^{\circ}\text{N}/299.2^{\circ}\text{E}$ ($A_{95} = 5.4^{\circ}$) with $N = 20$ dikes (Figure 7-10c; Table 7-3).

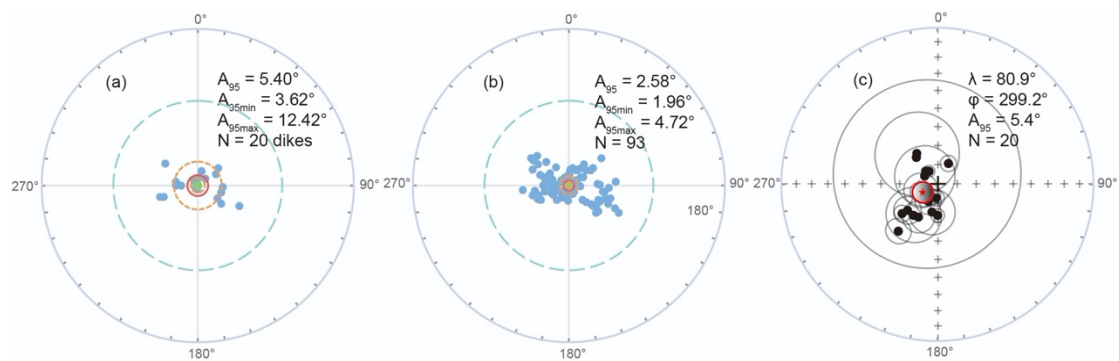


Figure 7-10. (a and b) Paleosecular variation of paleogeomagnetic field of our data set based on the Deenen et al. (2011) criteria at site-level and specimen-level, respectively; (c) Visual geomagnetic poles of each dikes and their mean-pole.

7.7 Summary

The first late Paleozoic - early Mesozoic paleomagnetic study for the Beishan Belt was carried out on the Early Permian doleritic dikes in the Gubaoquan area, southern Beishan.

(1) Thermomagnetic experiment and hysteresis loops revealed that magnetite seems the principal magnetic mineral with some pyrite, pyrrhotite and/or hematite of the doleritic dikes. AMS study showed no intensive deformation, and different magnetic fabrics between the mafic dikes and their country rocks.

(2) SEM observation indicated that irregular MD magnetite and titanomagnetite are main magnetic carriers and no authigenic mineral is found.

(3) The plagioclase collected from the margin of a wide doleritic dike gave $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages at 260-284 Ma.

(4) As supported by several lines of evidence, the magnetic remanence carried by the mafic dikes is probably primary with a remanence age at ~280 Ma. Moreover, a non/weak-tilting scenario is proposed based on field observation and AMS data. Accordingly, the first Early Permian paleomagnetic pole for the Beishan Belt is acquired at $\lambda/\phi = 80.9^\circ\text{N}/299.2^\circ\text{E}$ ($A_{95} = 5.4^\circ$) with $N = 20$ dikes.

Table 7-1. Sampling information in the Gubaoquan area, south Beishan

2017-10-01, we took paleomagnetic samples from Permian mafic dikes intruding into the orthogneisses of Beishan Complexe in Gubaoquan area, Beishan (N40°59'20.679", N95°02'9.964"---N40°59'37.295"N95°02'24.172")

NO	α	β	α_{sol}	T	Notes	α_{-corr}	$\Delta\alpha = -2.2$
Site BS30: dozens of mafic dikes intrude into orthogneisses of Beishan Complexe.							
MD01: N40°59'20.679", E95°02'9.964". A 2-3m mafic dike (50cm away from orthogneiss).							
BS30-01	243	62	287	11h02		240.4	-2.6
BS30-02	229	65	273	11h03		226.6	-2.4
BS30-03	236	60	280	11h05		234.1	-1.9
BS30-04	231	72	273	11h07	cc	227.6	-3.4
OG01: N40°59'20.605", E95°02'9.676". Orthogneisses (50cm away from mafic dike).							
BS30-05	274	43.5	315	11h18		272.5	-1.5
MD02: A 3m mafic dike.							
BS30-06	179	60	235	10h06		175.8	-3.2
BS30-07	159	57	197	11h22		155.6	-3.4
MD03: N40°59'20.635", E95°02'9.063". A 4m mafic dike.							
BS30-08	193	48	230	11h28	cc	190.3	-2.7
BS30-09	204	56	240	11h30		200.8	-3.2
MD04: N40°59'21.268", E95°02'8.323". A 1m mafic dike.							
BS30-10	226	53	263	11h34	cc	224.9	-1.1
BS30-11	236	56	272	11h36		234.5	-1.5
MD05: Separated from 30-11 by a 6m orthogneiss.							
BS30-12	255	43	287	11h46		252.4	-2.6
BS30-13	255	54	288	11h48		254	-1
OG02: N40°59'21.707", E95°02'7.543". 50m Orthogneisses. Samples 14,15 and 16 were collected from 50cm, 2m and 3m away from the contact.							
BS30-14	252	40	280	11h50		246.6	-5.4
BS30-15	212	49	240	11h58		209	-3
BS30-16	222	47	248	12h00		217.6	-4.4
MD06: N40°59'21.893", E95°02'5.441". A 4m mafic dike.							
BS30-17	220	45					
BS30-18	174	60					
BS30-19	154	65					
MD07: A 4m mafic dike separated from 30-19 by 6m orthogneisses.							
BS30-20	266	41					
BS30-21	239	50	265	12h15		239.3	0.3
MD08: A 3m mafic dike.							
BS30-22	207	48					
MD09: A 10m mafic dike.							
BS30-23	157	52	210	10h22		154.2	-2.8
BS30-24	156	47					
BS30-25	249	47					

BS30-26	167	58	219	10h23		163.5	-3.5
BS30-27	160	57	212	10h21	cc	156	-4
BS30-28	226	53	243	12h28		221.5	-4.5
BS30-29	244	65	260	12h30		239.2	-4.8
BS30-30	233	62	250	12h32		229.9	-3.1
BS30-31	230	51	246	12h34		226.5	-3.5
BS30-32	241	35	257	12h35		237.9	-3.1
OG03: orthogneisses, 1m away from mafic dike.							
BS30-33	250	65	236	12h38		???	
BS30-34	230	61	245	12h40	cc	227.6	-2.4
BS30-35	157	70					
BS30-36	207	48					
MD10: A 3m mafic dike.							
BS30-37	250	45					
BS30-38	235	62	286	10h32		232.5	-2.5
BS30-39	228	65	280	10h33		226.7	-1.3
MD11: A 3m mafic dike.							
BS30-40	252	67	301	10h35		248.2	-3.8
BS30-41	261	73	310	10h36		257.4	-3.6
BS30-42	204	62	252	10h37		199.6	-4.4
MD12: A 3m mafic dike.							
BS30-43	224	58					
BS30-44	246	82					
BS30-45	247	33					
MD13: A 3m mafic dike. Separated from 30-45 by 1m orthogneisses.							
BS30-46	254	75	304	10h39		252.1	-1.9
BS30-47	277	39	328	10h40		276.3	-0.7
BS30-48	260	53	338	10h41		???	
MD14: A 2m mafic dike. Separated from 30-48 by 3-5m orthogneisses.							
BS30-49	296	67	346	10h42		294.8	-1.2
BS30-50	245	74	292	10h44		241.3	-3.7
BS30-51	228	69					
MD15: A 2-3m mafic dike. Separated from 30-51 by 10m orthogneisses (including 3 felsic dikes).							
BS30-52	205	51					
BS30-53	208	52					
MD16: A 5m mafic dike. Separated from 30-53 by 1m orthogneisses.							
BS30-54	208	55					
BS30-55	196	80					
BS30-56	228	38	276	10h46		225.8	-2.2
BS30-57	263	49	310	10h48		260.3	-2.7

OG04: N40°59'25.065", E95°02'1.673". Orthogneisses. 58, 59, 60 were collected 1m, 2m, 5m away from contact.

BS30-58	315	42	3	10h49		313.5	-1.5
BS30-59	223	60	269	10h50		219.7	-3.3
BS30-60	262	47	308	10h51		259	-3

Site BS31: many mafic dikes intrude into orthogneisses of Beishan Complex.

MD17: N40°59'41.243", E95°02'25.908".

BS31-61	237	29	173	16h59		233.7	-3.3
BS31-62	168	58	106	16h57	CC	166.3	-1.7
BS31-63	120	68	59	16h56		119.1	-0.9
BS31-64	157	68	96	16h56		156.1	-0.9
BS31-65	61	73	358	16h55		57.9	-3.1
BS31-66	118	72	56	16h54	CC	115.7	-2.3
BS31-67	183	51	117	16h53		176.5	-6.5
BS31-68	83	53	20	16h53		79.5	-3.5

MD18: N40°59'38.506", E95°02'25.636".

BS31-69	142	48	81	16h52		140.2	-1.8
BS31-70	207	52	147	16h51		206	-1

MD19: N40°59'37.725", E95°02'24.556".

BS31-71	158	46	97	16h50		155.8	-2.2
BS31-72	150	77	90	16h49		148.6	-1.4
BS31-73	139	37	81	16h45		138.8	-0.2
BS31-74	140	59	81	16h44		138.5	-1.5

MD20: N40°59'37.333", E95°02'24.152".

BS31-75	129	63	70	16h42		127.1	-1.9
BS31-76	123	40			CC		

Note: 1. α , β , α_{sol} and T stand for the azimuth, hade, solar compass measurement and time of collection for samples, respectively; 2. α -corr means geographic azimuth, and $\Delta\alpha$ represents the angle difference between magnetic and geographic azimuths, and a mean $\Delta\alpha$ was calculated to correct the azimuth of samples without α_{sol} data; 3. MD means mafic dike and OG means orthogneisses (with yellow background).

Table 7-2. Plagioclase $^{40}\text{Ar}/^{39}\text{Ar}$ dating data of sample BS30-32 collected from the margin of doleritic dike MD09

Step #	$^{39}\text{Ar}_K$ (V)	$^{36}\text{Ar}_{\text{atm}}/^{39}\text{Ar}_K$ ($\pm 2\sigma$)	$^{37}\text{Ar}_{\text{Ca}}/^{39}\text{Ar}_K$ ($\pm 2\sigma$)	$^{38}\text{Ar}_{\text{Cl}}/^{39}\text{Ar}_K$ ($\pm 2\sigma$)	% $^{40}\text{Ar}^*$	$^{40}\text{Ar}^*/^{39}\text{Ar}_K$ ($\pm 2\sigma$)	Age (Ma) ($\pm 2\sigma$)
N216.MS1	BS30-32 (pl))	#11-B(AU)	J = .2761E-02 $\pm$.2862E-04				
# 1	-1.64E-08	-2.11E+00 $\pm$ 5.58E+00	-1.14E+00 $\pm$ 3.39E+02	1.00E+00 $\pm$ 2.31E+00	-114.4	332.50 $\pm$ 1314.60	1175.0 $\pm$ 3414
# 2	2.73E-07	1.60E+00 $\pm$ 3.46E-01	4.05E+00 $\pm$ 2.20E+01	4.35E-02 $\pm$ 1.29E-01	2.2	10.58 $\pm$ 71.60	51.9 $\pm$ 346
# 3	7.72E-07	1.84E+00 $\pm$ 2.71E-01	4.65E+00 $\pm$ 7.74E+00	-4.37E-02 $\pm$ 4.53E-02	6.2	36.21 $\pm$ 33.42	171.9 $\pm$ 151
# 4	1.58E-06	8.51E-01 $\pm$ 6.78E-02	1.16E+01 $\pm$ 4.05E+00	-5.43E-03 $\pm$ 2.81E-02	14.9	44.01 $\pm$ 13.36	206.9 $\pm$ 59
# 5	2.07E-06	4.28E-01 $\pm$ 3.92E-02	1.30E+01 $\pm$ 2.99E+00	-1.37E-02 $\pm$ 1.48E-02	28.2	49.60 $\pm$ 10.32	231.5 $\pm$ 45
# 6	3.84E-06	1.60E-01 $\pm$ 1.87E-02	1.39E+01 $\pm$ 2.63E+00	4.11E-03 $\pm$ 6.82E-03	43.0	35.75 $\pm$ 12.72	169.8 $\pm$ 58
# 7	1.41E-05	4.75E-02 $\pm$ 4.99E-03	9.36E+00 $\pm$ 6.49E-01	1.82E-03 $\pm$ 4.34E-03	77.4	48.00 $\pm$ 1.85	224.5 $\pm$ 8.1
# 8	9.90E-06	3.96E-02 $\pm$ 6.90E-03	9.35E+00 $\pm$ 7.82E-01	2.44E-03 $\pm$ 4.33E-03	79.8	46.23 $\pm$ 2.23	216.7 $\pm$ 9.8
# 9	1.87E-05	8.27E-02 $\pm$ 3.51E-03	7.06E+00 $\pm$ 5.48E-01	4.30E-03 $\pm$ 4.20E-03	68.8	54.00 $\pm$ 1.38	250.7 $\pm$ 6.0
# 10	1.12E-05	9.88E-02 $\pm$ 4.39E-03	6.17E+00 $\pm$ 1.04E+00	-8.29E-04 $\pm$ 2.80E-03	66.2	57.05 $\pm$ 1.92	263.9 $\pm$ 8.3
# 11	1.53E-05	1.32E-01 $\pm$ 4.40E-03	3.66E+00 $\pm$ 6.12E-01	-2.59E-03 $\pm$ 3.78E-03	59.9	57.94 $\pm$ 1.69	267.7 $\pm$ 7.2
# 12	1.75E-05	5.14E-02 $\pm$ 2.79E-03	4.38E+00 $\pm$ 6.63E-01	4.23E-04 $\pm$ 2.45E-03	79.8	60.09 $\pm$ 1.58	276.9 $\pm$ 6.7
# 13	1.77E-05	5.80E-02 $\pm$ 2.63E-03	5.03E+00 $\pm$ 5.14E-01	-1.13E-03 $\pm$ 1.82E-03	77.8	60.14 $\pm$ 1.57	277.1 $\pm$ 6.7
# 14	1.51E-05	5.64E-02 $\pm$ 2.57E-03	3.64E+00 $\pm$ 5.63E-01	2.91E-04 $\pm$ 2.91E-03	78.9	62.13 $\pm$ 1.55	285.6 $\pm$ 6.6
# 15	1.72E-05	3.84E-02 $\pm$ 2.23E-03	2.61E+00 $\pm$ 4.64E-01	1.15E-03 $\pm$ 3.25E-03	84.0	59.45 $\pm$ 1.89	274.1 $\pm$ 8.1
# 16	2.18E-05	3.02E-02 $\pm$ 1.98E-03	2.44E+00 $\pm$ 3.98E-01	2.17E-03 $\pm$ 2.24E-03	87.2	60.62 $\pm$ 1.34	279.1 $\pm$ 5.7
# 17	2.12E-05	4.05E-02 $\pm$ 1.75E-03	3.09E+00 $\pm$ 3.64E-01	2.29E-03 $\pm$ 2.30E-03	83.8	61.96 $\pm$ 1.41	284.8 $\pm$ 6.0
# 18	2.03E-05	3.88E-02 $\pm$ 2.01E-03	4.97E+00 $\pm$ 5.41E-01	2.32E-03 $\pm$ 2.53E-03	84.6	63.01 $\pm$ 1.09	289.3 $\pm$ 4.6
# 19	9.37E-06	2.83E-02 $\pm$ 1.59E-03	2.08E+00 $\pm$ 1.12E+00	2.20E-03 $\pm$ 4.29E-03	88.5	64.49 $\pm$ 1.47	295.6 $\pm$ 6.2
# 20	2.22E-05	5.69E-02 $\pm$ 2.47E-03	9.01E+00 $\pm$ 8.33E-01	7.89E-04 $\pm$ 2.21E-03	78.4	61.14 $\pm$ 1.34	281.4 $\pm$ 5.7
# 21	1.04E-05	4.90E-02 $\pm$ 2.80E-03	1.16E+01 $\pm$ 1.33E+00	6.31E-03 $\pm$ 4.01E-03	81.1	62.22 $\pm$ 1.37	285.9 $\pm$ 5.8
# 22	1.02E-05	4.66E-02 $\pm$ 2.97E-03	1.07E+01 $\pm$ 1.31E+00	5.89E-04 $\pm$ 3.26E-03	81.6	60.82 $\pm$ 2.26	280.0 $\pm$ 9.6
Total	2.61E-04	7.18E-02 $\pm$ 1.02E-03	5.83E+00 $\pm$ 1.79E-01	1.10E-03 $\pm$ 8.12E-04	73.4	58.40 $\pm$ 0.47	269.7 $\pm$ 5.6
N218.MS1	BS30-32 (pl))	#11-B(AU)	J = .2761E-02 $\pm$.2862E-04				

# 1	6.25E-06	3.41E-02	±	6.61E-03	3.95E+01	±	2.64E+00	3.29E-03	±	5.82E-03	85.0	57.05	±	3.39	263.8	±	14.6
# 2	1.03E-05	1.91E-02	±	3.90E-03	2.21E+01	±	1.56E+00	5.28E-04	±	3.51E-03	90.9	56.21	±	2.45	260.3	±	10.6
# 3	8.07E-06	1.62E-02	±	4.58E-03	1.45E+01	±	1.29E+00	1.07E-03	±	4.14E-03	92.4	57.99	±	2.65	267.9	±	11.4
# 4	1.07E-05	8.37E-03	±	3.36E-03	1.31E+01	±	1.08E+00	2.49E-04	±	4.09E-03	95.9	57.37	±	2.41	265.2	±	10.3
# 5	9.93E-06	1.29E-02	±	3.78E-03	1.19E+01	±	1.37E+00	1.78E-03	±	3.84E-03	93.7	56.61	±	2.35	262.0	±	10.1
# 6	6.90E-06	1.38E-02	±	5.17E-03	1.03E+01	±	1.50E+00	-8.47E-04	±	4.03E-03	93.2	55.45	±	3.07	257.0	±	13.3
# 7	5.91E-05	2.15E-02	±	9.78E-04	7.60E+00	±	2.37E-01	6.57E-04	±	9.52E-04	89.8	55.83	±	0.70	258.6	±	3.0
# 8	8.09E-06	4.40E-02	±	4.99E-03	2.32E+01	±	1.14E+00	2.99E-03	±	5.10E-03	82.4	60.70	±	2.05	279.5	±	8.8
# 9	1.00E-05	5.94E-02	±	4.67E-03	5.36E+01	±	2.06E+00	3.87E-03	±	4.04E-03	78.9	65.45	±	1.93	299.7	±	8.1
# 10	1.18E-05	4.11E-02	±	4.48E-03	6.81E+01	±	2.70E+00	-1.39E-03	±	2.89E-03	84.4	65.55	±	2.00	300.1	±	8.4
# 11	8.14E-07	5.01E-02	±	4.66E-02	5.87E+01	±	1.31E+01	2.43E-03	±	3.28E-02	81.9	67.12	±	17.78	306.7	±	75
Total	1.42E-04	2.53E-02	±	1.00E-03	2.08E+01	±	3.68E-01	9.63E-04	±	8.99E-04	88.6	58.02	±	0.56	268.0	±	5.7
N219.MS1	BS30-32 (pl))	#11-B(AU)	J = .2761E-02 ± .2862E-04														
# 1	3.48E-06	1.37E-01	±	1.46E-02	4.08E+01	±	4.17E+00	-4.20E-03	±	8.05E-03	58.7	57.47	±	5.01	265.7	±	22
# 2	3.85E-06	9.78E-02	±	1.17E-02	2.41E+01	±	2.50E+00	-2.77E-03	±	7.16E-03	65.7	55.28	±	4.31	256.2	±	18.6
# 3	5.31E-06	9.44E-02	±	8.12E-03	1.68E+01	±	1.54E+00	3.94E-03	±	6.22E-03	66.4	55.03	±	2.93	255.2	±	12.7
# 4	6.19E-06	1.31E-01	±	8.63E-03	1.78E+01	±	9.75E-01	-4.24E-03	±	5.71E-03	58.7	55.01	±	2.95	255.1	±	12.7
# 5	4.16E-06	1.25E-01	±	1.16E-02	1.27E+01	±	1.74E+00	-1.57E-03	±	1.06E-02	59.7	54.64	±	4.15	253.5	±	18.0
# 6	1.47E-05	8.39E-02	±	3.00E-03	1.10E+01	±	5.77E-01	1.02E-03	±	2.59E-03	69.8	57.24	±	1.29	264.7	±	5.6
# 7	8.88E-06	7.35E-02	±	5.48E-03	7.68E+00	±	1.21E+00	6.90E-03	±	5.98E-03	72.1	56.17	±	2.50	260.1	±	10.8
# 8	9.82E-06	5.01E-02	±	4.05E-03	9.86E+00	±	7.13E-01	4.18E-03	±	4.43E-03	79.8	58.37	±	1.81	269.5	±	7.8
# 9	2.45E-05	4.24E-02	±	2.38E-03	9.00E+00	±	4.64E-01	1.77E-03	±	1.86E-03	82.4	58.62	±	1.17	270.6	±	5.0
# 10	1.94E-05	7.10E-02	±	2.96E-03	2.23E+01	±	8.44E-01	7.15E-03	±	2.02E-03	75.6	64.98	±	1.55	297.7	±	6.6
# 11	1.74E-05	7.80E-02	±	3.16E-03	3.61E+01	±	1.22E+00	3.49E-03	±	3.42E-03	74.1	65.74	±	1.70	300.9	±	7.2
# 12	2.86E-06	1.12E-01	±	1.56E-02	2.57E+01	±	3.64E+00	-6.06E-03	±	7.76E-03	66.1	64.75	±	6.44	296.7	±	27
Total	1.21E-04	7.59E-02	±	1.38E-03	1.80E+01	±	3.19E-01	2.54E-03	±	1.16E-03	72.7	59.83	±	0.62	275.8	±	5.9

The Ar-Ar dating experiments were conducted in the Ar-Ar laboratory of ISTO (Institut des Sciences de la Terre d'Orléans) at Orléans University.

Table 7-3. Paleomagnetic results of the doleritic and orthogneissic specimens from the Gubaoquan area, south Beishan.

Site	S-Lat (N)	S-Long (E)	n/n ₀ (N)	R/N	D _g (°)	I _g (°)	k	α ₉₅ (°)	P-Lat (N°)	P-Long (E°)	A ₉₅ (°)
MD01	40°59'20.679"	95°02'9.964"	5/6	5/0	196.9	-48.7	1644.8	1.9	72.2	219.3	2.0
OG01	40°59'20.605"	95°02'9.676"	1/1	1/0	163.6	-25.4	2.5	12.7			
MD02			2/2	2/0	176.9	-51.1	948.5	8.1	80.5	291.1	9.0
MD03	40°59'20.635"	95°02'9.063"	3/3	3/0	163.3	-43.7	71.2	14.7	69.2	322.0	14.4
MD04	40°59'21.268"	95°02'8.323"	2/2	2/0	199.2	-48.7	149.3	20.6	70.7	215.0	22.0
MD05			5/5	5/0	170.7	-58.4	523.9	3.3	82.6	353.1	4.2
OG02	40°59'21.707"	95°02'7.543"	1/4	0/1	125.1	7.9	5.7	26.2			
MD06	40°59'21.893"	95°02'5.441"	5/5	5/0	161.6	-46.5	90.0	8.1	69.9	329.5	8.3
MD07			2/2	2/0	169.6	-54.1	1881.7	5.8	79.6	330.5	6.8
MD08			2/2	2/0	159.2	-56.9	8184.0	2.8	73.6	359.5	3.5
MD09			11/12	11/0	188.6	-56.1	354.8	2.4	82.0	215.2	3.0
OG03			1/7	1/0	157.9	-29.0	8.1	37.5			
MD10			6/6	6/0	188.6	-54.4	1270.4	1.9	80.9	224.1	2.3
MD11			6/7	6/0	194.7	-64.6	534.6	2.9	78.1	152.7	4.2
MD12			4/6	4/0	156.8	-29.0	129.8	8.1	57.5	320.0	6.6
MD13			5/5	5/0	165.6	-35.9	240.3	4.9	65.6	309.5	4.3
MD14			6/6	6/0	166.7	-40.5	75.9	7.7	68.9	311.0	7.2
MD15			2/2	2/0	185.6	-53.8	344.3	13.5	82.0	239.6	15.8
MD16			3/3	3/0	162.4	-54.7	150.4	10.1	75.0	348.0	12.0
OG04	40°59'25.065"	95°02'1.673"	0/4								
MD17	40°59'41.243"	95°02'25.908"	13/15	13/0	176.3	-49.9	94.2	4.3	79.3	292.3	4.7
MD18	40°59'38.506"	95°02'25.636"	3/3	3/0	173.7	-55.8	761.8	4.5	83.2	323.6	5.4
MD19	40°59'37.725"	95°02'24.556"	5/6	5/0	176.8	-54.7	961.3	2.5	83.7	299.6	3.0
MD20	40°59'37.333"	95°02'24.152"	2/2	2/0	187.5	-54.5	36.5	42.6	81.6	227.9	50.5
Mean (MD)			20/20	20/0	174.9	-51.4	47.4	4.8	80.9	299.2	5.4

Abbreviations: Slat (N°)/ Slong (N°), latitude and longitude of sampling sites. n/n₀ (N), number of specimens (sites) in calculation/ totally treated.

D_g and I_g, declination (D) and inclination (I) in geological (g) coordinates; k, precision parameter; α₉₅, 95% confidence interval.

Plat_g, Plong_g, and A₉₅, latitude and longitude of paleomagnetic poles in geographic (g) coordinates, and 95% confidence interval.

MD, mafic dike, and OG, orthogneisses (with yellow background).

Chapter 8. Discussion

8.1 Implications for the late Paleozoic post-orogenic kinematic evolution of the SW CAO B through the Early Permian Yili paleomagnetic pole

8.1.1 Comparisons of paleomagnetic poles between the Yili Block and the adjacent blocks

Until about ten years ago, no paleomagnetic data were documented from the Yili Block. Some results were obtained from Permo-Carboniferous rocks in the western Kyrgyz Tianshan (Bazhenov et al., 1993, 1999, 2003), and most of the samples were collected from the orogenic belt where local rotations took place along major faults within the tectonic belt, therefore they cannot represent the paleopoles the Yili Block (Bazhenov et al., 1999; Wang et al., 2007). Conducting the first paleomagnetic study on the sedimentary and volcanic rocks from the Yili Block, Wang et al. (2007) provided Late Carboniferous and Late Permian paleomagnetic poles for the Yili Block. In the study presented in this thesis, we obtain an Early Permian pole that may provide more data for unravelling the late Paleozoic kinematic evolution of this block with respect to the neighboring ones (Table 8-1).

Table 8-1 presents a synthesis of paleomagnetic poles of Late Carboniferous (C₂), Early Permian (P₁) and Late Permian (P₂) from the Tarim, Yili and South Junggar blocks. Among the 5 Late Permian poles of the South Junggar Block, the former three are consistent (Table 8-1; Sharps et al., 1992; Choulet et al., 2011). The fourth one comes from Nie et al. (1993) and is significantly different from the former three. Choulet et al. (2013) conducted a paleomagnetic study basically in the same area as Nie et al. (1993) and acquired another Late Permian paleomagnetic pole which is indistinguishable from that of Nie et al. (1993) with an angular difference of $4.0 \pm 20.4^\circ$. Based on regional geology and stratigraphic observations, Choulet et al. (2013)

considered that the studied strata could have underwent a local rotation during syn-tectonic sedimentation. Accordingly, the latter two poles have been removed from the mean pole calculation, and the mean P_2 pole for the South Junggar Block is recalculated at 71.1°N , 6.1°E , $A_{95} = 14.5^\circ$ (Table 8-1). It is also worth noting that the P_1 pole for the South Junggar may be questionable as it was acquired from only 4 sites.

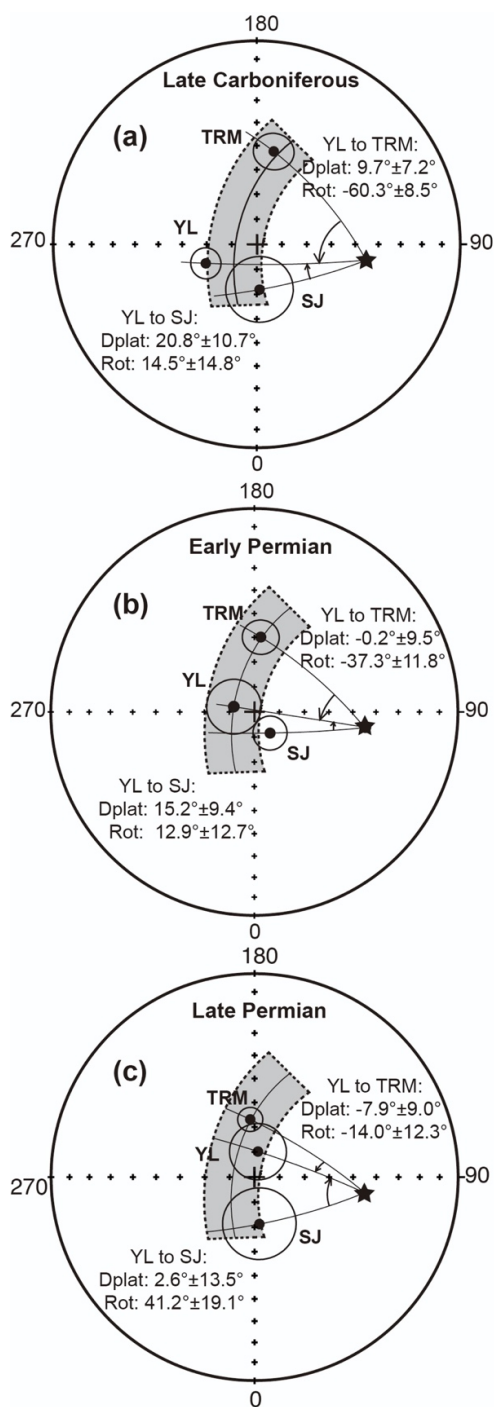


Figure 8-1. Equal-area projections of paleomagnetic poles of the Yili (YL), Tarim (TRM), and South Junggar (SJ) blocks in (a) Late Carboniferous, (b) Early Permian, and (c) Late Permian. Dplat and Rot stand for convergence and relative rotation (with corresponding uncertainties), respectively. The positive (or negative) mean values stand for latitudinal shortening (or extension) or clockwise (counterclockwise) rotation, respectively.

Figure 8-1 shows the mean paleomagnetic poles of the Yili, Tarim and South Junggar blocks for the C_2 , P_1 and P_2 periods. The distribution of these poles seems quite dispersed with significant angular differences until the Late Permian. For the Late Carboniferous, the pole of the Yili Block is $52.3^\circ \pm 7.2^\circ$ and $23.0^\circ \pm 10.7^\circ$ away from those of the Tarim and South Junggar blocks. The angular differences

between the Yili poles and the poles of the Tarim and South Junggar blocks progressively decrease to $29.7^{\circ} \pm 9.5^{\circ}$ and $17.2^{\circ} \pm 9.4^{\circ}$ for the Early Permian, then to $12.5^{\circ} \pm 9.0^{\circ}$ and $28.2^{\circ} \pm 13.5^{\circ}$ for the Late Permian, respectively (Table 8-2).

However, the above mentioned paleomagnetic poles are mainly distributed along small circles centered at the sampling sites. In other words, these polar angular differences may essentially correspond to the relative rotations and the movements along block boundaries, while the crustal shortening among these blocks is relatively weak or negligible. The relative rotations of the Yili Block with respect to the Tarim and South Junggar blocks are calculated as $-60.3^{\circ} \pm 8.5^{\circ}$ and $14.5^{\circ} \pm 14.8^{\circ}$, $-37.3^{\circ} \pm 11.8^{\circ}$ and $12.9^{\circ} \pm 12.7^{\circ}$, and $-14.0^{\circ} \pm 12.3^{\circ}$ and $41.2^{\circ} \pm 19.1^{\circ}$ since C₂, P₁, and P₂, respectively (Figure 8-1). The latitudinal movements of the Yili Block with respect to the Tarim and South Junggar blocks are estimated as $9.7^{\circ} \pm 7.2^{\circ}$ and $20.8^{\circ} \pm 10.7^{\circ}$, $-0.2^{\circ} \pm 9.5^{\circ}$ and $15.2^{\circ} \pm 9.4^{\circ}$, and $-7.9^{\circ} \pm 9.0^{\circ}$ and $2.6^{\circ} \pm 13.5^{\circ}$ since C₂, P₁, and P₂, respectively (Figure 8-1; Table 8-2).

8.1.2 New quantitative constraints on the late Paleozoic kinematic evolution of the SW CAOB and its tectonic implications

Wang et al. (2007) conducted the first quantitative estimation on the relative movements among the three major blocks in the SW CAOB (the Yili, Tarim and Junggar blocks). Due to limited data from the Yili and Junggar blocks, they grouped the paleomagnetic poles for the Late Carboniferous and the Late Permian stages and concluded that these blocks shared similar latitudes but experienced differential rotations during a single C₂-P₂ period. With newly obtained Paleozoic and Mesozoic paleomagnetic poles from the south and west Junggar areas, Choulet et al. (2011) argued that the Junggar domain cannot be considered as a rigid block at least until the end of the Paleozoic, and only the South Junggar experienced a similar tectonic evolution with the Yili and Tarim blocks in the late Paleozoic. Our new P₁

paleomagnetic results for the Yili Block allows us to depict a more detailed kinematic history among these blocks.

According to our and the previous paleomagnetic results, the relative movements between these major blocks are indicated by changes of angular differences between their paleomagnetic poles with the time. For the Yili and Tarim blocks which are separated by the Nikolaiev Tectonic Line (NTL; e.g. Burtman, 1975) and the Nalati Fault (Wang et al., 2008, 2010; Gao et al., 2009), a significant rotation with a weak latitudinal movement occurred during C₂ to P₁, and an equally important rotation but no obvious latitudinal movement happened from P₁ to P₂ (Figure 8-1; Table 8-2). The Yili Block shows a counter-clockwise (CCW) rotation of $23.0^\circ \pm 11.6^\circ$ and a northward movement of $9.9^\circ \pm 9.5^\circ$ with respect to the Tarim Block during the C₂-P₁ period. These results are quite consistent with the dextral kinematics along the NTL and Nalati Fault (e.g. Yin and Nie, 1996; Wang et al., 2007, 2008, 2010). The timing of rotation is also in agreement with the available ⁴⁰Ar/³⁹Ar age obtained along the NTL and Nalati Fault (286.4 ± 2.4 Ma; Yin and Nie, 1996; Laurent-Charvet et al., 2003). Wang et al. (2007) and Choulet et al. (2011) suggested that the relative rotations were accommodated by large movements along the major ductile shear zones, and the Euler pole of this relative movement is located to the north of these blocks. Thus, the relative rotation between these two blocks could be accommodated by an eastward extrusion of the Yili Block with respect to the Tarim Block. Moreover, as the NTL and Nalati Fault strike NE-SW, the weak northward movement of the Yili Block with respect to the Tarim Block might occur as a simple ‘sliding’ along the fault instead of a crustal shortening. Assuming that the NTL and Nalati Fault kept the same shape during C₂-P₁ as that described in Wang et al. (2007) for C₂-P₂, the Euler pole documented by Wang et al. (2007) can be adopted to estimate the movement along these faults. With a radius of about 1440 km and a rotation angle of $23.0^\circ \pm 11.6^\circ$, the lateral displacement during C₂-P₁ (300-270 Ma) is calculated at 580 ± 290 km, corresponding to a displacement rate of $\sim 19.3 \pm 9.7$ mm/yr. (Figures 8-2a-b).

Since then, the relative rotation between two blocks continued ($-23.3^\circ \pm 13.6^\circ$) from P_1 to P_2 but the latitudinal movement became negligible ($7.7^\circ \pm 10.5^\circ$). This is consistent with the dextral strike-slip faulting along the Nalati Fault that was still active at least up to 250 Ma as suggested by $^{40}\text{Ar}/^{39}\text{Ar}$ dating results (de Jong et al., 2009; Wang et al., 2010). Similarly, using a Euler pole radius of 1440 km and a rotation angle of $23.3^\circ \pm 13.6^\circ$, the lateral displacement during P_1 - P_2 (270-250 Ma) along this fault can be estimated at 585 ± 340 km. The corresponding displacement rate is calculated as $\sim 29.3 \pm 17.0$ mm/yr. (Figures 8-2b-c).

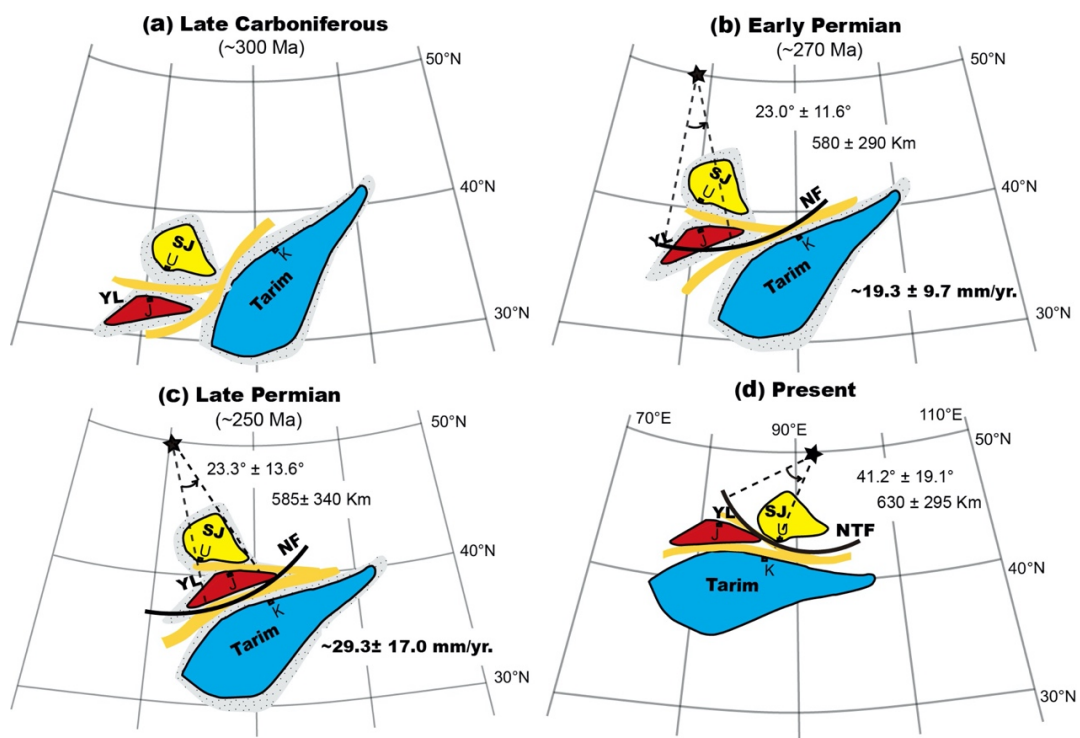


Figure 8-2. Tentative paleogeographic reconstruction maps of the Yili, Tarim and South Junggar blocks for (a) Late Carboniferous, (b) Early Permian, (c) Late Permian, and (d) Present with the relative movements. Shaded zones around each block indicate the loss of continental margin during Cenozoic intra-continental deformation. J, K, and U stand for the positions of Jinghe, Korla, and Urumqi cities, respectively.

The Yili and South Junggar blocks are separated by the North Tianshan Fault (NTF) (e.g. Zhou et al., 2001; Zhao et al., 2003; Wang et al., 2006, 2007, 2009) that extends eastwards and merges into the Main Tianshan Shear Zone (MTSZ) (Laurent-

Charvet et al., 2002, 2003). Dextral sense of movement was observed along these faults (Laurent-Charvet et al., 2002, 2003; Wang et al., 2006, 2007, 2009). $^{40}\text{Ar}/^{39}\text{Ar}$ dating constrained the deformation ages from 290 Ma to 240 Ma (Yin and Nie, 1996; Zhou et al., 2001; Laurent-Charvet et al., 2002, 2003; de Jong et al., 2009). In the same way, we calculated the relative latitudinal movement and rotation between Yili and South Junggar during C_2 to P_1 period as $5.6^\circ \pm 11.4^\circ$ and $1.6^\circ \pm 15.6^\circ$, and during the P_1 to P_2 as $12.6^\circ \pm 13.2^\circ$ and $-28.3^\circ \pm 18.3^\circ$, respectively. The tectonic interpretation of these calculations must be cautiously made because the paleomagnetic data from South Junggar are still scarce, and the quality of current poles needs to be reinforced. Keeping the large uncertainties in mind, no significant change of angular differences between the C_2 poles and P_1 poles of two blocks can be recognized (Figures 8-1a-b), indicating that no relative movement occurred between two blocks during C_2 to P_1 , likely because both blocks moved with similar directions and magnitudes. By contrast, angular differences between the P_2 poles of two blocks are obviously distinct from the angular differences of their P_1 poles, and a significant counterclockwise rotation of $-28.3^\circ \pm 18.3^\circ$ of the Yili Block with respect to the South Junggar Block can be calculated (Figures 8-1b-c and Figures 8-2b-c). Such remarkable relative rotation was probably resulted from obviously different motions both in directions and in magnitudes of two blocks during P_1 to P_2 period.

Furthermore, our analysis reveals that the Yili Block experienced a large rotation of $41.2^\circ \pm 19.1^\circ$ with respect to South Junggar along the North Tianshan Fault after the Late Permian. This large relative rotation corresponds to a lateral displacement of 630 ± 295 km on the basis of the Euler pole defined by Choulet et al. (2011) with a radius of 880 km (Figures 8-2c-d). This result confirms the paleomagnetic study by Choulet et al. (2011) showing that the dextral shearing continued after the Late Permian, and is consistent with the dextral motion of the North Tianshan Fault whose timing is dated at the Early Triassic by $^{40}\text{Ar}/^{39}\text{Ar}$ method (244.7 ± 2.6 Ma; Laurent-Charvet et al., 2003).

8.2 Constraining the Intracontinental Tectonics of the SW Central Asian Orogenic Belt by the Early Permian Paleomagnetic Pole for the Turfan-Hami Block

8.2.1 Synthesis of compiled paleomagnetic poles for major blocks in the SW CAO

In this Ph.D. study, we acquired the first Early Permian paleomagnetic pole for the Turpan-Hami Block (#1 in Figure 8-3a, Tables 8-3 and 8-4). The only Early Permian paleomagnetic pole for the Yili Block was published by Zhu et al. (2018) (#2 in Figure 8-3a, Tables 8-3 and 8-4). For the South Junggar block, only one Early Permian Paleomagnetic pole acquired from limited data set (4 sites) by Choulet et al. (2011) was published, and it is quite different from the Late Carboniferous and the Late Permian South Junggar poles (Table 8-3; Li et al., 1991; Sharps et al., 1992; Nie et al., 1993; Choulet et al., 2011, 2013). Since the reliability of this pole is questionable (Zhu et al., 2018), we prefer to use a Late Carboniferous - Late Permian mean pole in further comparisons (#3 in Figure 8-3a, Tables 8-3 and 8-4). The Early Permian paleomagnetic pole for the Tarim Block are recalculated through six former poles (#4 in Figure 8-3a, Tables 8-3 and 8-4; Bai et al., 1987; Sharps et al., 1989; Li et al., 1995; Fang et al., 1996).

The data from Cogné et al. (1995) and Huang et al. (2004) were combined and recalculated to acquire a mean Cretaceous pole for the Tu-Ha Block (#5 in Figure 8-3b, Tables 8-3 and 8-4). As these data were all obtained from sedimentary rocks, they could suffer from inclination shallowing effect which is frequently reported in the Central Asia (Gilder et al., 2003; Tan et al., 2003). E/I correction was therefore performed. Although limited site-mean directions ($n = 24$) were used in this correction, the E-W flat VGPs distribution indicated a significant inclination shallowing (Tauxe et al., 2004) and the flattening factor was calculated as 0.58. Accordingly, an inclination-corrected paleomagnetic pole was acquired (#6 in Figure 8-3b, Tables 8-3 and 8-4). No Cretaceous pole was reported for the Yili Block up to now. As for the South Junggar

block, Chen et al. (1991) acquired the only Cretaceous paleomagnetic pole of the South Junggar block (#7 in Figure 8-3b, Tables 8-3 and 8-4). E/I correction was also applied on this data set as it was obtained from sedimentary rocks. No obvious inclination shallowing was found, which is supported by the circularly distributed VGPs. More reliable mean Cretaceous poles were acquired for the Tarim block where the paleomagnetic research degree is quite high (#8 in Figure 8-3b, Tables 8-3 and 8-4; Li et al., 1988; Chen et al., 1992; Li et al., 1995; Fang et al., 1997, 2001; Meng et al., 1998; Gilder et al., 2003).

8.2.2 Kinematic constraints on the intracontinental tectonics of the SW CAO

With the aim of a better understanding of the kinematic evolution among major blocks (the Yili, South Junggar, Tarim and Turpan-Hami blocks) in Central Asia, we compiled published paleomagnetic poles and re-calculated the mean poles for these blocks (Table 8-3; see details in section 8.2.1). The Early Permian pole of the Turpan-Hami Block that we report is significantly different from those of South Junggar and Tarim, with angular differences of $22.0^{\circ} \pm 7.2^{\circ}$ and $35.0^{\circ} \pm 5.9^{\circ}$, respectively. However, the Early Permian poles of the Turpan-Hami and Yili blocks are undistinguishable with an angular difference of $6.6^{\circ} \pm 9.0^{\circ}$ (Figure 8-3a; Table 8-4). Cogné et al. (1995) and Huang et al. (2004a) have published Cretaceous paleomagnetic results for the Turpan-Hami Block. Taking into account the possible inclination shallowing of their data, the Cretaceous paleomagnetic pole of the Turpan-Hami Block is nearly undistinguishable from results from the Junggar and Tarim blocks with angular differences of $1.7^{\circ} \pm 5.1^{\circ}$ and $5.7^{\circ} \pm 5.1^{\circ}$, respectively (Figure 8-3b; Table 8-4).

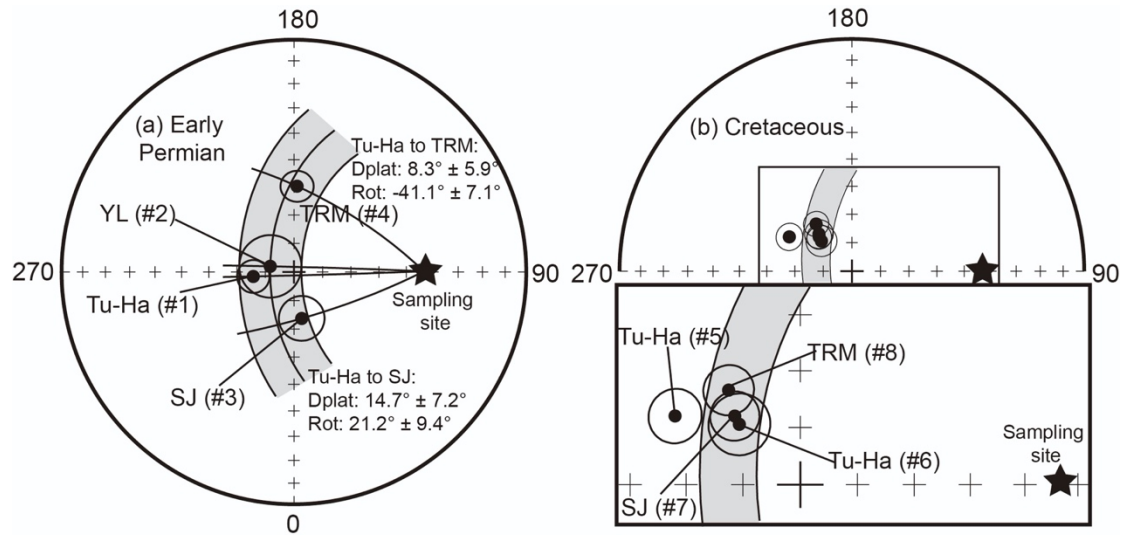


Figure 8-3. Equal-area projections of paleomagnetic poles of the Turpan-Hami (Tu-Ha), Yili (YL), South Junggar (SJ) and Tarim (TRM) blocks in (a) the Early Permian and (b) Cretaceous. Reference numbers (#No.) stand for different paleomagnetic poles listed in Tables 8-3 and 8-4, which are consistent. Relative movements between different paleomagnetic poles are shown in Table 8-4.

Table 8-4. Relative movements of the Turpan-Hami block with respect to the surrounding blocks.

Period	Blocks	Angular difference (°)	Plat ± Δplat (°)	Rot ± Δrot (°)
Early Permian - present	TH (#1) vs. YL (#2)	6.6 ± 9.0	6.1 ± 9.0	-4.2 ± 10.9
	TH (#1) vs. SJ (#3)	22.0 ± 7.2	14.7 ± 7.2	21.2 ± 9.4
	TH (#1) vs. TRM (#4)	35.0 ± 5.9	8.3 ± 5.9	-41.1 ± 7.1
Cretaceous - present	TH (#6) vs. SJ (#7)	1.7 ± 5.1	-1.0 ± 5.1	-1.6 ± 6.0
	TH (#6) vs. TRM (#8)	5.7 ± 5.1	-2.9 ± 5.1	-6.5 ± 6.0
	SJ (#7) vs. TRM (#8)	4.1 ± 4.5	-1.9 ± 4.5	-5.0 ± 5.2

Abbreviations: TH, YL, SJ and TRM stand for the Turfan-Hami, Yili, South Junggar and Tarim blocks. Plat ± Δplat and Rot ± Δrot correspond to relative latitudinal displacements and rotations (and the errors) between blocks.

The reference numbers #no. of each paleomagnetic pole are consistent with that in Figure 8-3 and Table 8-3. The relative movements between the blocks are calculated using an average sampling site at: 43.4°N, 90.3°E.

Our study provides the first Early Permian paleomagnetic pole for the Turpan-Hami Block, allowing us to discuss late Paleozoic paleogeographic position of this block and its movement with respect to surrounding crustal elements. The insignificant

angular difference of $6.6^{\circ} \pm 9.0^{\circ}$ between the Early Permian paleomagnetic poles of Turpan-Hami and Yili blocks corresponds to relative latitudinal movement of $6.1^{\circ} \pm 9.0^{\circ}$ and relative rotation of $-4.2^{\circ} \pm 10.9^{\circ}$ (Table 8-4), which are unrecognizable within errors, suggesting that these two blocks have likely maintained approximate relative positions since the Early Permian.

Nevertheless, the observed large angular difference of $35.0^{\circ} \pm 5.9^{\circ}$ between the Early Permian paleomagnetic poles of Turpan-Hami and Tarim (Table 8-4) indicates significant relative movements between these blocks since the Early Permian. Such significant relative movement is essentially represented by a counterclockwise rotation of $41.1^{\circ} \pm 7.1^{\circ}$ of the Turpan-Hami Block with respect to Tarim and by a relatively small latitudinal movement of $8.3^{\circ} \pm 5.9^{\circ}$ (Table 8-4). Similar large counterclockwise rotations were also documented for the Yili Block with respect to Tarim from the Late Carboniferous to the Late Permian, and were interpreted as the result of the large-scale dextral strike-slip movements between the Yili and Tarim blocks (Wang et al., 2007; Zhu et al., 2018). In fact, large-scale strike-slip shearing and related transpression and/or transtension were documented to be the principal pattern of the intracontinental tectonics in the SW CAOB (Allen et al., 1993; Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2009, 2014; Choulet et al., 2011, 2013; Zhu et al., 2018). Considering that the Turpan-Hami and Yili blocks may have kept their relative position since the Early Permian, these two blocks could have subsequently experienced a similar kinematic evolution. Therefore, this counterclockwise relative rotation of about 40° between the Turpan-Hami Block and Tarim can also be accommodated by large-scale right-lateral strike-slip movements along the NF-MTSZ. This hypothesis is supported by (1) structural and kinematic studies which show that ductile shearing along the MTSZ has been of dextral sense (Yin and Nie, 1996; Laurent-Charvet et al., 2002, 2003; Wang et al., 2007, 2014), and (2) multi-mineral $^{40}\text{Ar}/^{39}\text{Ar}$ dating on mylonitic rocks from the MTSZ yields cooling ages ranging from 290-245 Ma (Laurent-Charvet et al., 2003; de Jong et al., 2009). As for the relatively small latitudinal difference of $8.3^{\circ} \pm 5.9^{\circ}$

between the Turpan-Hami and Tarim blocks, it can be considered as the result of a northward relative movement of the Turpan-Hami Block together with the Yili Block with respect to the Tarim Block that might have occurred as a simple ‘sliding’ along the NE-SW-striking NF-MTSZ, instead of crustal shortening (Zhu et al., 2018).

Between the Turpan-Hami and South Junggar blocks, significant relative movements since the Early Permian are suggested by an angular difference of $22.0^{\circ} \pm 7.2^{\circ}$ between their Early Permian paleomagnetic poles. Such discordance can be explained by a clockwise relative rotation of $21.2^{\circ} \pm 9.4^{\circ}$ and northward latitudinal movement of $14.7^{\circ} \pm 7.2^{\circ}$ of the Turpan-Hami Block with respect to the South Junggar Block since the Early Permian. According to previous paleomagnetic studies (Choulet et al., 2011; Zhu et al., 2018), large relative rotation corresponding to a dextral displacement of 630 ± 295 km along the NTF occurred between the Yili and South Junggar blocks after the Late Permian. Considering the aforementioned possibly similar kinematic evolution between the Turpan-Hami and Yili blocks, the clockwise rotation of the Turpan-Hami Block with respect to the South Junggar Block may also reflect large-scale lateral displacement between two blocks along the NTF. The latitudinal movement of $14.7^{\circ} \pm 7.2^{\circ}$ between the Turpan-Hami and South Junggar blocks possibly sheds light on a late Paleozoic intra-arc basin in the Bogda Belt and is discussed below.

8.2.3 Tectonic implications for a late Paleozoic intra-arc basin in the Bogda Belt

Being located between the Turpan-Hami and South Junggar blocks, the Bogda Belt experienced Early Permian volcanism and marine sedimentation, Late Permian to Triassic terrigenous deposition, Cretaceous uplift and Cenozoic shortening (Hendrix et al., 1992; Carroll et al., 1995; Greene et al., 2001, 2005; Wartes et al., 2002; Shu et al., 2011; Chen et al., 2015; Xie et al., 2016). Thus, pulses of late Paleozoic to Cenozoic significant crustal uplift and related shortening may account for the latitudinal movement since the Early Permian between the Turpan-Hami and South Junggar blocks. Numerous studies have estimated the magnitude of N-S shortening since the Cretaceous

across the Tianshan Belt. Cenozoic crustal deformation in the western Chinese Tianshan Belt corresponds to ~150 km shortening between the Tarim and Junggar basins (Avouac et al., 1993; Avouac and Tapponnier, 1993). Facies analysis of the Tadjik basin indicates a total crustal shortening of 300 km within the Tianshan and Pamir ranges (Burtman, 2000). In addition, Chen et al. (1991) proposed that the crustal shortening between the Junggar and Tarim blocks since the Cretaceous is 150-175 km in the eastern part of the Tianshan Range, and reaches ~280 km in its western part (Kashgar). These conclusions are generally in agreement with each other and also consistent with our estimates based on integrated Cretaceous paleomagnetic results that show estimated relative latitudinal movements of 1.0° - 2.9° since the Cretaceous (Figure 8-3b; Table 8-4). Clearly, these values are much smaller than the total relative latitudinal movements since the Early Permian, which should, therefore, be the result of Early Permian to Cretaceous tectonism.

Greene et al. (2001, 2005) argued for an Early Jurassic initial uplift of the Bogda Belt based on petrographic approaches and trace-element and isotopic geochemistry, which are also supported by paleocurrent, sedimentary facies, and other stratigraphic evidence (Hendrix et al., 1992; Wartes et al., 2002). Unconformities and evolution of sedimentary compositions in the Turpan-Hami basin both suggest a continuous uplift and erosion of the Bogda Belt in the Mesozoic (Greene et al., 2001). Though the Mesozoic uplift of the Bogda Belt may contribute to the crustal shortening, its magnitude is difficult to quantitatively evaluate due to the insufficiency of data. Mesozoic paleomagnetic poles of the Turpan-Hami and Junggar blocks are therefore badly needed to assess this issue. Nevertheless, a preliminary estimate could be carried out based on Early Triassic paleomagnetic poles of South Junggar (Choulet et al., 2013) and Tarim (Zhu et al., 1998). Relative latitudinal movement between the South Junggar and Tarim blocks since the Early Triassic is accordingly approximated as $5.0^{\circ} \pm 12.2^{\circ}$. In paleomagnetism, when uncertainty is large, the calculated value generally provides nothing more than rough qualitative information. Therefore, several hundred kilometers

of crustal shortening should have occurred between these two blocks since the Early Triassic. Furthermore, the crustal shortening that took place in the Triassic and Jurassic might be of similar or smaller magnitude than that which occurred since the Cretaceous, as the great India-Eurasia collision began in the Cenozoic. Consequently, the crustal shortening of 300-350 km between South Junggar and Tarim blocks, and smaller (~300 km) between South Junggar and Turpan-Hami blocks since the Early Triassic is permitted. The abovementioned crustal shortening between the Turpan-Hami and South Junggar blocks is still much smaller than our estimate of 1630 ± 800 km ($14.7^\circ \pm 7.2^\circ$) since the Early Permian. Therefore, pre-Triassic tectonic event(s) should also be included in this important crustal shortening.

The late Paleozoic tectonic evolution of the Bogda Belt has been extensively studied, mainly based on geochemistry and geochronology of magmatic rocks. Different tectonic settings have been proposed, such as an island arc (Carroll et al., 1995; Xie et al., 2016), a continental rift (Shu et al., 2011; Xia et al., 2008, 2012), a mantle plume-related large igneous province (Su et al., 2011; Xia et al., 2008), and a post-collision extensional belt (Chen et al., 2011; Ma et al., 2015). Recently, Zhang et al. (2017) suggested that upper Carboniferous volcanic rocks in the Dashitou area from the eastern Bogda Belt formed in an extensional regime associated with a subduction-related environment and proposed an intra-arc rifting tectonic setting. Wali et al. (2018) compiled literature data and suggested that the Carboniferous to Early Permian magmatic rocks in the Bogda Belt generally show consistent arc-type features with increasing mantle input through time, consistent with upwelling of asthenospheric mantle under an intra-arc rifting or pull-apart extension in supra-subduction zone (Wang et al., 2009; Branquet et al., 2012). In addition, Shu et al. (2011) reported an Early Permian deepening sedimentary environment in the Baiyanggou section, southwest of the Bogda Belt, characterized by olistostrome, chert-bearing turbidite, bimodal volcanic rocks including pillow basalts, likely caused by initiation of intra-arc rifting. The felsic volcanic rocks yield zircon U-Pb ages of 311 - 297 Ma (Shu et al.,

2011; Xie et al., 2016). Therefore, an intra-arc rift basin likely existed in the Bogda Belt during late Carboniferous to early Permian times. The closure of this intra-arc rift basin was poorly constrained as no typical subduction or collision-related records were recognized. However, several lines of evidence indicate important crustal uplift and shortening posterior to the Early Permian. These arguments include: (1) a regional unconformity of upper Permian with a thick molasse sequence, (2) a change of sedimentary environments from marine to terrigenous, and (3) tight folds and pervasive cleavage of the Carboniferous to Lower Permian metasedimentary rocks in the Bogda area (XBGMR, 1993; Shu et al., 2011; Wali et al., 2018). Therefore, we propose that the Bogda intra-arc rift basin was likely closed during the Middle to Late Permian time.

Obviously, the closure of this intra-arc basin should be another important contribution to the relative latitudinal movement between the Turpan-Hami and South Junggar blocks since the Early Permian. Keeping the large uncertainties in mind, by subtracting the maximum estimated amount of crustal shortening between these two blocks since the Early Triassic (~300 km), a northward movement of at least 530 km for the Turpan-Hami Block with respect to South Junggar could be proposed, which resulted from the closure of the intra-arc rift basin in the Bogda Belt. Clearly, more paleomagnetic studies and further geologic investigations are needed to verify this hypothesis.

8.3 Preliminary geological implications from the Baluntai migmatite

Since the Baluntai migmatites probably underwent significant horizontal-axis rotation, it is unsuitable to use migmatites to constrain the paleolatitude of the Baluntai Block (Central Tianshan). Moreover, our former paleomagnetic studies on the Carboniferous sandstones were also unable to give a constraint. Therefore, we suggest that even though the paleogeographic position of the Central Tianshan is of great geological importance, robust paleomagnetic data are challenging to obtain for this block. We appeal for more paleomagnetic studies on no/low-degree

metamorphic/deformed sedimentary rocks instead of the migmatites in the Central Tianshan.

Although the paleolatitude of the Baluntai Block is unknown, the magnetic declination may still be meaningful, and thus reflecting the relative rotations between different blocks. If our remanence-acquisition model is correct, then the $\sim 30^\circ$ rotation of the Baluntai migmatite along a horizontal axis with NW-SE strike (regional tectonic line, present coordinates) barely affect the magnetic declination (the northward magnetic declination is quasi-perpendicular to the rotation axis). Thus, our magnetic declination could give first-order constraints on the relative rotations of the Baluntai Block with respect to the other major blocks.

No matter in the geological coordinates or the tectonic coordinates, at the first-order, the Baluntai Block shows similar declination with respect to the Tarim Block; in contrast, it experienced large-scale clockwise rotations with respect to the Yili-Turpan-Hami block and especially the South Junggar Block (Figure 8-4).

These observations suggest significant dextral strike-slip displacements along the Nalati Fault and North Tianshan-Middle Tianshan Shear zone. Concerning the relative rotation between the Baluntai and Tarim, due to the large uncertainty, we are unable to discuss more details. Nevertheless, this conclusion confirms again the importance of dextral strike-slip tectonics as the major pattern of the late Paleozoic evolution in the SW CAO.

In the latest Carboniferous, the closure of the Ural Ocean marks the welding among Laurussia Block, Siberia Craton, and Kazakhstan continent. However, the eastward propagation of the deformation was ongoing. Due to its special crustal structure, of different terranes including microcontinents, island arcs, subduction wedges, etc., the deformation in the CAO was heterogeneously distributed. As a result, deformation was localized in pre-existing weak zones (e.g. ancient sutures, continental faults, etc.) and kilometers-scale dextral strike-slip faults developed around microcontinents. The Central Tianshan coupled with its northern and southern dextral strike-slip fault (NTF-

MTSZ, NF-BF) is an ideal example of this tectonic pattern.

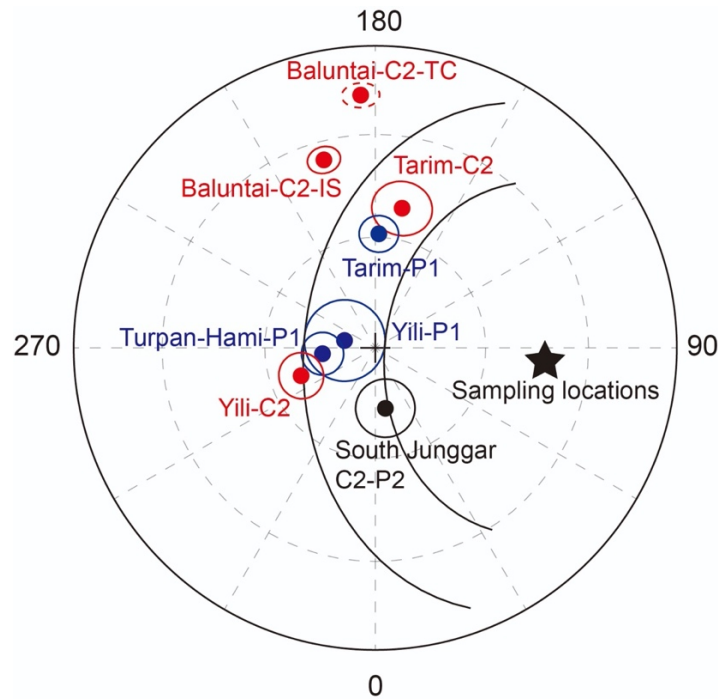


Figure 8-4. Compilation of Late Carboniferous to Early Permian Paleomagnetic poles of the major blocks in the SW CAOB. Note: due to the scarcity of paleomagnetic data for the South Junggar, we use a Late Carboniferous- Late Permian mean pole.

Furthermore, since the geological background of the Baluntai Block is significantly different from the North Tianshan and South Tianshan and two large-scale strike-slip faults developed at its north and south boundary, it is also possible that the Central Tianshan had a low paleolatitude in the Moscovian but experienced large-magnitude dextral strike-slip displacement since then. However, the $\sim 19.4^\circ$ (~ 2100 km) latitudinal difference cannot be totally accommodated by the dextral strike-slip displacement, thus, we propose that strike-slip displacements and foliation tilting both contributed to such a low paleolatitude of the Baluntai Block.

8.4 Tectonic implications from the Early Permian Beishan paleomagnetic “Bridge”

8.4.1 Compilation of the Early Permian paleomagnetic poles for the major blocks in the southern CAO

Previous published Early Permian paleomagnetic poles for the major blocks in the South CAO were compiled in Table 8-5 and Figure 8-5.

In the middle part of the south CAO, i.e., the ‘bridge’ between the western and eastern CAO, the Early Permian paleomagnetic poles of Beishan from this study and Dunhuang from Xu et al. (2019) were compiled.

In the western part of the south CAO, the Early Permian paleomagnetic poles for the Yili and Turpan-Hami blocks were provided by Zhu et al. (2018, 2019). For the south Junggar, only one Early Permian paleomagnetic work has been published with only 4 sites by Choulet et al. (2011); therefore, we also compiled available paleomagnetic data from Late Carboniferous to Late Permian (Li et al., 1991; Sharps et al., 1992; Choulet et al., 2011) and calculated a mean pole to better represent the possible Early Permian pole for the south Junggar (Zhu et al., 2019). As for the Tarim block, abundant works have been carried out and the results are quite consistent. Hence, we compiled together the data cited in Zhu et al. (2019), the results from Cheng et al. (1983), and a recently acquired high-quality result reported by Wei et al. (2020) to calculate a mean paleomagnetic pole for the Tarim block at the Early Permian.

In the eastern part of the south CAO, 11 paleomagnetic poles were compiled. For the Alax block, only one pole was reported by Wu et al. (1993); Concerning the North China Block, three poles were compiled spanning the latest Late Carboniferous to Middle Permian (Ma et al., 1992; Li et al., 2012; Zhang et al., 2018). These poles are consistent in paleolatitude but with significant relative rotation, therefore, we have not calculated a mean pole for the North China Craton. The situation is the same when

it comes to the paleomagnetic poles for the Mongolia block. Seven poles covering the Late Carboniferous to Late Permian were compiled (Chen et al., 1997; Pruner, 1992; Xu et al., 1997; Li et al., 2012; Zhao et al., 2013, 2020; Zhang et al., 2018). However, the Early Permian poles for the Mongolia block proposed by Pruner (1992) and Xu et al. (1997) are of small site numbers and large uncertainties. Based on its equatorial paleolatitude and nearly perpendicular magnetic declination comparing with other results, Zhao et al. (2020) suggested that the area where Zhang et al. (2018) conducted their work probably underwent significant posterior. Therefore, we exclude those three poles from our compilation (Figure 8-5a).

Notably, due to the consistent paleolatitudes and distinct magnetic declinations of the poles for the North China Block and Mongolia Block, we use only the Early Permian poles reported by Ma et al. (1992) and Zhao et al. (2013) to perform paleogeographic reconstruction of the eastern CAOB.

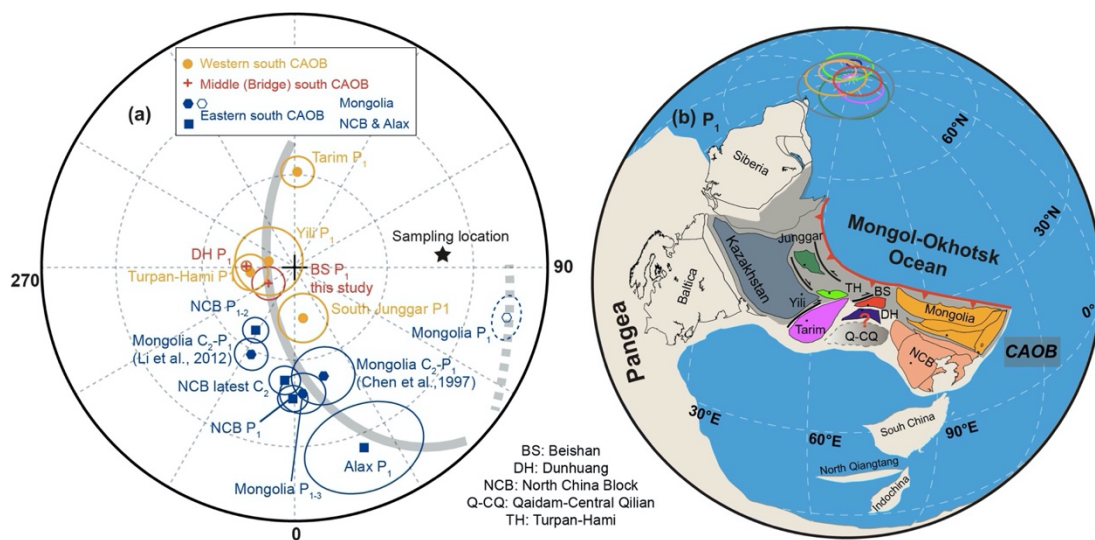


Figure 8-5. (a) Compilation of Early Permian paleomagnetic poles for major blocks in the southern CAOB (see Table 8-5; Ma et al., 1992; Chen et al., 1997; Li et al., 2012; Zhao et al., 2013; Zhang et al., 2018; Wei et al., 2020). (b) Early Permian paleogeographic reconstruction of south CAOB.

8.4.2 Tectonic implications

According to this Ph.D. study, a latitudinal difference between Beishan and Dunhuang blocks is estimated as $7.0^{\circ} \pm 4.5^{\circ}$. However, no paleomagnetically significant latitudinal movement is suggested between them since the Early Permian, because: 1) the confidence interval ($A_{95} = 1.6^{\circ}$ for 22 sites) of the Dunhuang pole (Xu et al., 2019), retrieved from sandstones, sedimentary tuffs, and basalts, seems too small to represent paleosecular variation according to statistical criteria defined by Tauxe and Kent (2004) and Deenen et al. (2011); 2) the magnetic inclination shallowing in Xu et al. (2019) should be treated more carefully, especially when mean inclination of basalt sites (only two) is 5.3° shallower than that of sedimentary sites (Zhu et al., 2019); and 3) a possible weak tilting ($<15^{\circ}$) of dikes could result in a magnetic inclination shallowing of up to $\sim 6^{\circ}$ for Beishan.

Therefore, we argue that the existence of an oceanic basin between Beishan and Dunhuang block in the Early Permian is not supported by paleomagnetic data. This conclusion is in agreement with previous arguments: 1) Su et al. (2013) indicated that the P_1 mafic-ultramafic complexes in Beishan have no similarity with Alaskan-type (subduction-related) intrusive; Chen et al. (2016) reported the P_1 peperites in the Liuyuan area, indicating that the local subaqueous basaltic lava flows are not dismembered ophiolitic components; 2) Widespread P_1 alkali-rich high-potassium granites (Zhang et al., 2012; Li et al., 2013; Xue et al., 2018; Yuan et al., 2018) and bimodal igneous series (Su et al., 2011) in Beishan support a P_1 post-collisional extensional setting; 3) Niu et al. (2018) presented the Carboniferous-Permian marine rift sequences in the Beishan region, and proposed the local closure of the PAO before the Carboniferous.

From a larger scale, taking in mind a possible weak tilting, the P_1 pole of Beishan is close to that of Yili, Turpan-Hami, and Dunhuang (Figure 8-5a), suggesting that these blocks have been amalgamated together at least since the P_1 and no paleomagnetically significant relative movement occurred among them thereafter. In

contrast, the P_1 poles of Tarim and South Junggar are obviously different from those of the above blocks with significant relative rotation and weak latitudinal movement (Figure 8-5a). These movements were previously studied and interpreted as the result of the large-magnitude strike-slip displacements along mega-shear zones and the closure of pull-apart basins triggered by strike-slip displacements since the P_1 (Wang et al., 2007, 2009, 2014; Zhu et al., 2018, 2019).

According to available data from the southern CAO, the following phenomena can be observed (Figure 8-5a): there are no obvious latitudinal differences among major blocks either in the Tianshan or Beishan or Mongolia belts since the P_1 ; the P_1 paleomagnetic poles from the major blocks in the southern CAO distribute along a small circle, revealing large relative rotations among them since then; the P_1 paleolatitude of the major blocks in the southern CAO decreases eastward (reference site: 41.0°N , 95.0°E) from $\sim 34^\circ\text{N}$ (west), $\sim 29^\circ\text{N}$ (middle) to $\sim 28^\circ\text{N}$ (east). Combining the previous P_1 paleomagnetic results, our newly obtained pole for Beishan can be considered as a ‘bridge’ between the SW and SE CAO that provides the following important constraints on the transition from convergent to intracontinental tectonics in the southern CAO: 1) the PAO was likely closed at least in the P_1 ; thereafter large-magnitude strike-slip displacements and related local vertical-axis rotations (Zhao et al., 2015) acted as the major intracontinental tectonics; 2) the major blocks in the southern CAO presented a northwest-southeast distributed paleogeography in the P_1 (Figure 8-5b). Presently, the highest latitude of the Mongolia Belt within the southern CAO is mainly due to the Mesozoic closure of the Mongol-Okhotsk Ocean (Zhao et al., 2020).

Table 8-1. Compilations of Late Carboniferous to Late Permian paleomagnetic poles from the Yili, Tarim and South Junggar blocks

<i>Blocks</i>	<i>Areas</i>	<i>Ages</i>	<i>N</i>	<i>S-Lat (°N)</i>	<i>S-Long (°E)</i>	<i>P-Lat (°N)</i>	<i>P-Long (°E)</i>	<i>A₉₅ (°)</i>	<i>References</i>
Yili	Mean ^a	C ₂	15			68.6	290.6	6.1	Wang et al. (2007)
	Mean ^b	P ₁	11			81.5	256.5	10.9	This study
	Gongliu	P ₂	5	43.4	82.5	79.7	172.0	11.3	Wang et al. (2007)
Tarim	Mean ^c	C ₂				51.5	169.1	7.7	Wang et al. (2007)
	Mean ^d	P ₁				59.7	175.1	7.1	Zhao et al. (2014)
	Mean ^e	P ₂				66.8	184.1	4.9	Wang et al. (2007)
South	Urumqi	C ₂	15	43.8	87.8	71.9	4.7	13.3	Li et al. (1991)
Junggar	Urumqi	P ₁	4	43.8	87.8	79.5	36.6	6.8	Choulet et al. (2011)
	Urumqi	P ₂	4	43.8	87.8	75.0	13.3	22.1	Sharps et al. (1992)
	Urumqi ^f	P ₂	78s	43.8	87.8	77.7	0.5	5.5	Sharps et al. (1992)
	Urumqi	P ₂	6	43.8	87.8	60.4	4.7	5.4	Choulet et al. (2011)
	Tianzhi ^g	P ₂	7	44.0	88.1	76.0	193.0	16.2	Nie et al. (1993)
	Tianchi ^g	P ₂	6	43.9	88.1	80.0	191.6	12.4	Choulet et al. (2013)
	Mean	P ₂	3			71.1	6.1	14.5	This study

Abbreviations: *S-Lat/S-Long*, latitude/longitude of sampling sites.

P-Lat/P-Long, latitude/longitude of paleomagnetic poles.

A₉₅ is the radius with which mean directions lies within 95% confidence.

N is the number of sites.

^a Mean of poles obtained from the Zhaosu and Xinyuan areas, southern Yili Block (Wang et al., 2007).

^b Mean of poles obtained from the Jinghe and Nileke areas, northern and central Yili Block (This study).

^c Mean of poles from Sharps et al. (1992), Van Der Voo (1993), Nie et al. (1993), Gilder et al. (1996), Zhao et al. (2003), and Harkand et al. (2005).

^d Mean of poles from Bai et al. (1987), Sharps et al. (1989), Gilder et al. (1996), and Fang et al. (1998).

^e Mean of poles from Li et al. (1989), Li et al. (1991), Meng, (1991), and Gilder et al. (1996).

^f Pole calculated from 52 out of 78 samples analyzed with great circle method.

^g poles not used for calculating the P₂ mean pole of the South Junggar. More details see Discussion section 7.1.1.

Table 8-2. Relative movements of the Yili Block with respect to the surrounding blocks during the late Paleozoic

Period	Blocks	Angle difference (°)	Plat ± Δplat (°)	ROT ± ΔROT (°)
C ₂	YL vs. TRM	52.3 ± 7.2	9.7 ± 7.2	-60.3 ± 8.5
	YL vs. SJ	23.0 ± 10.7	20.8 ± 10.7	14.5 ± 14.8
P ₁	Yl vs. TRM	29.7 ± 9.5	-0.2 ± 9.5	-37.3 ± 11.8
	YL vs. SJ	17.2 ± 9.4	15.2 ± 9.4	12.9 ± 12.7
P ₂	YL vs. TRM	12.5 ± 9.0	-7.9 ± 9.0	-14.0 ± 12.3
	YLvs. SJ	28.2 ± 13.5	2.6 ± 13.5	41.2 ± 19.1

Abbreviations: YL, SJ and TRM stand for the Yili Block, South Junggar Block and Tarim Block, respectively.

Plat ± Δplat and ROT±ΔROT correspond to relative latitudinal displacement and rotation (and their errors) between blocks.

Note that the relative movements between the blocks are computed using an average sampling site at: 44.2°N, 82.0°E.

Table 8-3. Compilation of the Early Permian and the Cretaceous paleomagnetic poles for major blocks in the SW CAOB

Blocks	Locality	Ages	Lithology	Slat	Slong	N	Plat	Plong	A ₉₅ (dp/dm)	# Ref.	Reference
Tu-Ha basin	Qijiaojing	P ₁	Sandstones	43.4	91.7	7	75.7	276.3	5.7	#1	This study
	Daheyang		Sandstones	43.3	88.9	3					
	Shengjingkou	J ₃ -K ₁	Sandstones	43.0	90.0	13	65.1	223.8	5.1/8.4		Cogne et al. (1995)
	Shengjingkou	K ₁	Sandstones	42.9	89.6	10	64.5	250.0	5.6		Huang et al. (2004)
	Tuyugou		Sandstones	42.9	89.7	3					
	Mean	K		43.0	89.8	24	64.9	241.2	4.7	#5	this study
	Mean-corrected	K		43.0	89.8	24	75.0	225.2	5.4	#6	this study
Yili	Jinghe	P ₁	Basalts, sandstones, limestones	44.5	81.6	4	81.5	256.5	10.9	#2	Zhu et al. (2018)
	Nileke			43.7	82.3	7					
South Junggar	Urumqi ^a	C ₂	Mafic dikes	43.8	87.8	7	54.6	173.5	14.0		Li et al. (1991)
	Urumqi ^a	C ₂	Andesitic flows, tuffs, interbedded clastics	43.8	87.8	4	73.4	96.5	27.4		Li et al. (1991)
	Urumqi	C ₂	Sedimentary rocks, volcanoclastics	43.8	87.8	15	71.9	4.7	13.3		Li et al. (1991)
	Urumqi	P ₁	Sandstones, siltstones	43.8	87.8	4	79.5	36.6	6.8		Choulet et al. (2011)
	Urumqi	P ₂	Sandstones, siltstones	43.8	87.8	6	60.4	4.7	5.4		Choulet et al. (2011)
	Urumqi	P ₂	Andesitic tuffs, siltstones,	43.8	87.7	78s	77.7	0.4	5.5		Sharps et al. (1992)
	Urumqi	P ₂	Mudstones	43.8	87.7	4	75.2	13.4	22.1		Sharps et al. (1992)
	Tianchi ^b	P ₂	Sandstones, siltstones	44.0	88.1	7	76.0	199.5	16.2		Nie et al. (1993)
	Tianchi ^b	P ₂	Sandstones	43.9	88.1	6	80.0	191.6	12.4		Choulet et al. (2013)
	Mean	C ₂ -P ₂				5P	73.5	9.5	7.9	#3	Choulet et al. (2011) and this study
	Manas	J ₃ -K ₁	Sandstones	44.2	86.0	13	72.3	227.3	4.8		Chen et al. (1991)
	Manas	K ₂ -E ₁	Sandstones	44.2	86.0	9	74.3	223.1	6.4		Chen et al. (1991)
	Mean	K				22	73.4	223.6	4.3	#7	Chen et al. (1991)
Tarim	Keping	P ₁	Sandstones, basalts, clastics			22	56.1	179.4	4.4		Fang et al. (1996)

Kuergan	P ₁	Sandstones, andesites	41.4	79.9	14	63.2	171.4	7.0/9.9	Li et al. (1995)
Mazha	P ₁	Diabase, andesites	36.5	77.0	8	60.6	188.0	13.9/23.2	Li et al. (1995)
Subasi	P ₁	Limestones	40.5	78.8	11	65.1	162.5	4.1	Sharps et al. (1989)
Yingan & Camp 66	P ₁	Basalts	40.7	79.6	21	54.5	172.3	4.0	Sharps et al. (1989)
Sishichang of Akesu	P ₁	Sandstones	40.8	79.8	26	56.5	190.1	4.6	Bai et al. (1987)
Mean	P ₁				6P	59.7	177.9	5.6	#4 Zhao et al. (2014) and this study
Paicheng	J ₃ -K ₁	Sandstones, mudstones	41.8	82.0	6	64.7	209.0	8.6	Li et al. (1988)
Tuoyun, SW Tarim	K ₁	Basalts, red beds	40.2	75.3	13	64.1	172.1	12.0	Gilder et al. (2003)
Tuoyun, SW Tarim ^c	K ₁	Red beds	40.2	75.3	5	53.2	183.4	6.1	Gilder et al. (2003)
Kuqa	K ₁	Mudstones, siltstones	41.6	83.5	18	66.2	234.7	5.1/8.8	Fang et al. (2001)
Tuoyun, SW Tarim	K ₁	Basalts	40.2	75.3	9	69.7	211.6	12.5	Meng et al. (1998)
Yingjisha, Kashgar	K ₁	Sandstones, mudstones	38.5	76.4	3	70.4	212.1	6.6/10.8	Chen et al. (1992)
Wuqia, Kashgar	K ₁	Sandstones, mudstones	39.5	75.0	7	66.3	226.6	9.0/15.9	Chen et al. (1992)
Kuqa river of Kuqa	K ₂	Mudstones	41.6	83.5	10	64.0	229.0	11.0	Fang et al. (1997)
Tuoyun, SW Tarim	K ₂	Basalts	40.2	75.4	8	77.1	211.9	11.3/17.0	Li et al. (1995)
Wuqia, Kashgar	K ₂	Sandstones, mudstones	39.5	75.0	11	70.8	222.6	5.4/8.9	Chen et al. (1992)
Yingjisha, Kashgar	K ₂	Sandstones, mudstones	38.5	76.4	6	71.0	234.0	6.8/11.6	Chen et al. (1992)
Kuqa	K ₂	Mudstones, sandstones	41.6	83.5	7	66.3	224.1	8.3	Li et al. (1988)
Mean	K				11P	69.2	217.1	4.4	#8 this study

Abbreviations: Slat (Plat), the latitude of the sampling site (paleomagnetic pole); Slong (Plong), the longitude of sampling site (paleomagnetic pole).

A95, 95% confidence interval; dp/dm, the two axes of an oval of confidence with 95%.

N is the number of sites (specimens, s; poles, P).

Ref., reference number labeled in Figure 8-3 and Table 8-4. Details are introduced in Text S1.

^a Discard from mean pole calculation due to important uncertainties on the bedding surface (Li et al., 1991). ^b Discard from mean pole calculation due to possible local rotations of studied strata (Choulet et al., 2013). ^c Discard from mean pole calculation due to inclination shallowing (Gilder et al., 2003).

Table 8-5. Compilation of the Early Permian paleomagnetic poles for major blocks in the south Central Asian Orogenic Belt (reference location: 41.0° N, 95.0° E)

Blocks	Locality	Ages	Lithology	Slat (°N)	Slong (° E)	N	Plat (°N)	Plong (° E)	A ₉₅ (dp/dm) (°)	Paleolat (°N)	References
Middle ('Bridge') South Central Aisan Orogenic Belt											
Beishan	Gubaoquan	P ₁	doleritic dikes	41.0	95.0	20	80.9	299.2	5.4	32.6 ± 5.4	this study
Dunhuang	South of Dushan	P ₁	Sandstones. sedimentary tuffs. basalts	40.6	94.1	22	74.5	268.5	1.6	25.6 ± 1.6	Xu et al. (2019)
Western South Central Asian Orogenic Belt											
Yili	Jinghe	P ₁	Basalts. sandstones.	44.5	81.6	4	81.5	256.5	10.9	32.9 ± 10.9	Zhu et al. (2018)
	Nileke		limestones	43.7	82.3	7					
Turpan-Hami	Qijiaoqing	P ₁	Sandstones	43.4	91.7	7	75.7	276.3	5.7	26.7 ± 5.7	Zhu et al. (2019)
	Daheyang		Sandstones	43.3	88.9	3					
South Junggar	Urumqi	C ₂	Sedimentary rocks. volcaniclastics	43.8	87.8	15	71.9	4.7	13.3		*Li et al. (1991)
	Urumqi	P ₁	Sandstones. siltstones	43.8	87.8	4	79.5	36.6	6.8		*Choulet et al. (2011)
	Urumqi	P ₂	Sandstones. siltstones	43.8	87.8	6	60.4	4.7	5.4		*Choulet et al. (2011)
	Urumqi	P ₂	Andesitic tuffs.	43.8	87.7	78s	77.7	0.4	5.5		*Sharps et al. (1992)
	Urumqi	P ₂	siltstones. Mudstones	43.8	87.7	4	75.2	13.4	22.1		*Sharps et al. (1992)
	Mean	C ₂ -P ₂				5P	73.5	9.5	7.9	40.2 ± 7.9	in Zhu et al. (2019)
Tarim	Keping	P ₁	Basalts. sandstones	40.8	79.8	24/290s	62.8	191.5	3.2		Cheng et al. (1983)
	Sishichang of Akesu	P ₁	Basalts. sandstones	40.8	79.8	26/317s	56.5	190.1	4.6		*Bai et al. (1987)
	Subasi	P ₁	Limestones	40.5	78.8	11/67s	65.1	162.5	4.1		*Sharps et al. (1989)
	Yingan & Camp 66	P ₁	Basalts	40.7	79.6	21/145s	54.5	172.3	4.0		*Sharps et al. (1989)
	Kuergan	P ₁	Sandstones. andesites	41.4	79.9	14	63.2	171.4	7.0/9.9 (8.3)		*Li et al. (1995)
	Mazha	P ₁	Diabase. andesites	36.5	77.0	8	60.6	188.0	13.9/23.2 (18.0)		*Li et al. (1995)
	Keping	P ₁	Sandstones. basalts. clastics			22	56.1	179.4	4.4		*Fang et al. (1996)

	Keping	P ₁	Basalts. sandstones. limestones. tuffs	40.7	79.6	51/413s	50.0	170.7	3.6		Wei et al. (2020)
	Mean	P ₁					59.0	178.2	5.1	37.5 ± 5.1	this study
Eastern South Central Asian Orogenic Belt											
Alax	Zhongwei/ Shandan/ Hangwula	P ₁	Gray/red sandstone	37.6	101.3	5/22s	23.9	21.2	12.8/21.4 (16.6)	27.3 ± 16.6	Wu et al. (1993)
North China	Baiyinwula	latest C ₂	tuffaceous sandstone	42.5	119.3	4	53.1	355	4.8	26.5 ± 4.8	Li et al. (2012)
	Ordos Basin	P ₁	Siltstone. sandy mudstone	35.5	110.3	6	47.0	359.2	3.1/5.7 (4.2)	25.3 ± 4.2	Ma et al. (1992)
	Huade	P ₁₋₂	basaltic andesite. lava. tuff	42.1	114.2	24	66.2	328.1	4.1	24.7 ± 4.1	Zhang et al. (2018)
Mongolia	Inner Mongolia	C ₂ -P ₁	Tuffs. tuffaceous sandstones. siltstones	45.5	117.2	5	53.4	14.8	9.7	37.1 ± 9.7	Chen et al. (1997)
	Inner Mongolia	C ₂ -P ₁	crystal tuffs	44.3	113.3	7	58.5	333.4	5.2	20.7 ± 5.2	Li et al. (2012)
	^a Inner Mongolia	P ₁	basalts. andestes. and tuff	43.6	117.3	14	41.6	246.1	7.5	-3.4 ± 7.5	Zhang et al. (2018)
	Inner Mongolia	P ₁₋₃	basalts, tuffs, and sandstones	44.3	118.4	6	48.7	3.7	5.3/9.1 (6.9)	28.8 ± 6.9	Zhao et al. (2013)
	Mongolia	P ₁	rhyolites, dacites	49.1	113.4	7	-14.9	76.8	5.7	31.6 ± 5.7	Zhao et al. (2020)
	Mongolia	P ₁		47.8	107.1	4	44.8	335.1	11.6	11.3 ± 11.6	Pruner (1992)
	Russia	P ₁		51.5	115.4	5	33.8	207.8	26.3	7.0 ± 26.3	Xu et al. (1997)

Abbreviations: Slat (Plat), the latitude of the sampling site (paleomagnetic pole); Slong (Plong), the longitude of sampling site (paleomagnetic pole).

A95, 95% confidence interval; dp/dm, the two axes of an oval of confidence with 95% confidence. N is the number of sites (speimens, s; poles, P).

*These references have been cited and introduced in our previous study. More information could be found in Zhu et al. (2019).

^a This paleomagnetic pole is significantly different from others, and its magnetic declination is nearly perpendicular to others', which could result from local deformation.

Mongolian data with yellow background are excluded from Figure 8-5a due to their samll site numbers and large uncertainties.

Chapter 9. Conclusions and Perspectives

9.1 Conclusions

As discussed in Chapter 1, despite decades of investigations, numerous fundamental issues remain hotly debated in the geoscience community regarding the CAOBT tectonic evolution. One main topic of this research project was to enlarge the scope of prior studies by combining several approaches (structural field-work, paleomagnetism, geochronology) across a vast, still poorly studied but critical, area extending from Yili in the west to Beishan in the east. Integrating available results with these new data, the following main conclusion can be reached.

1. The first Early Permian paleomagnetic poles for the Yili Block ($\lambda = 81.5^\circ\text{N}$, $\phi = 256.5^\circ\text{E}$, $A_{95} = 10.9^\circ$ with $N = 11$ sites), the Turpan-Hami Block ($\lambda = 75.7^\circ\text{N}$, $\phi = 276.3^\circ\text{E}$, $A_{95} = 5.7^\circ$ with $N = 10$ sites), and the Beishan Belt ($\lambda = 80.9^\circ\text{N}$, $\phi = 299.2^\circ\text{E}$, $A_{95} = 5.4^\circ$ with $N = 20$ dikes) were obtained, shedding light on the late Paleozoic paleogeographic configuration and the tectonic transition from collision/accretion to intracontinental tectonics of the southern CAOBT. Although no plausible paleomagnetic pole could be provided, our multidisciplinary study of the Baluntai Block (Chinese Central Tianshan) allows a preliminary understanding of different acquisition mechanisms between the magnetic fabric and magnetic remanence of the Baluntai migmatite and how these are presumably related to the tectonic evolution;

2. New quantitative constraints on the late Paleozoic kinematic evolution of the SW CAOBT are acquired by comparing the Early Permian Yili paleomagnetic pole with those published for the Yili, Tarim, and South Junggar blocks. Between the Yili and Tarim blocks, significant relative movements took place along major strike-slip faults during the Late Carboniferous to Early Permian (580 ± 290 km) and the Early to Late Permian (585 ± 340 km), and the displacement rate increased from the Late Carboniferous to Early Permian ($\sim 19.3 \pm 9.7$ mm/yr) to the Early to Late Permian

($\sim 29.3 \pm 17.0$ mm/yr). The relative movement is paleomagnetically negligible between the Yili and South Junggar blocks during the Late Carboniferous to Early Permian, but a significant relative rotation of $28.3^\circ \pm 18.3^\circ$ in the Late Permian and a lateral displacement of 630 ± 295 km after the Late Permian occurred between these two blocks;

3. The Turpan-Hami Block experienced relative rotations of $-41.1^\circ \pm 7.1^\circ$ and $21.2^\circ \pm 9.4^\circ$ with respect to the Tarim and South Junggar blocks since the Early Permian, respectively. They correspond to large-magnitude strike-slip displacements along major strike-slip faults, characterizing the main intracontinental tectonic features of the SW CAOB since the late Paleozoic. Furthermore, small but significant latitudinal movements of $8.3^\circ \pm 5.9^\circ$ and $14.7^\circ \pm 7.2^\circ$ are revealed for the Turpan-Hami Block with respect to the Tarim and South Junggar blocks, respectively. The former could be interpreted as the result of a strike-slip displacement along the NE-SW-striking Nalati Fault-Middle Tianshan Shear Zone, instead of crustal shortening. The latter was possibly accommodated by the closure of a ~ 530 km-width intra-arc rift basin that existed between the South Junggar and Turpan-Hami blocks in the Early Permian and by subsequent uplifts of the Bogda Range;

4. The migmatization in the Baluntai area occurred at ~ 377 Ma and probably continued to ~ 310 Ma, which likely resulted from diachronous closure of the back-arc basin(s) between the Central Tianshan and the South Tianshan. At ~ 310 Ma, the Baluntai migmatites underwent a fast uplift/exhumation. The migmatite preserves a “primary” horizontal magnetic fabric acquired during the migmatization, indicating the convergence and amalgamation of the Central Tianshan and South Tianshan terminated in the Moscovian (ca. 312-306 Ma). Paleomagnetic results combined with presently available data indicate that the ChRM was likely acquired after a folding with unknown attitude (probably with a NW-SE strike) and underwent a significant rotation thereafter. Hence, the paleolatitude of the Baluntai Block is unknown. However, the magnetic declination may still be meaningful, showing obvious clockwise relative rotation with

respect to Yili-Turpan-Hami and South Junggar, and suggesting significant dextral strike-slip displacements along the Nalati Fault and North Tianshan-Middle Tianshan Shear zone since the latest Carboniferous;

5. The latitudinal difference between Beishan and Dunhuang since the Early Permian is calculated as $7.0^{\circ} \pm 4.5^{\circ}$. However, this “paleomagnetically significant” latitudinal movement is neither supported by geological evidence nor reanalysis of published paleomagnetic data. Therefore, the middle segment of the Paleo-Asian Ocean should have closed at least in the Early Permian;

6. Since the Early Permian, no obvious latitudinal differences have been observed among the paleomagnetic poles of major blocks either in the Tianshan or Beishan or Mongolia belts; the paleomagnetic poles of the major blocks in the southern CAOB distribute along a small circle, revealing large relative rotations among them since then. The Early Permian paleolatitude of the major blocks in the southern CAOB decreases eastwards (reference site: 41.0°N , 95.0°E) from $\sim 34^{\circ}\text{N}$ (west), $\sim 29^{\circ}\text{N}$ (middle) to $\sim 28^{\circ}\text{N}$ (east). Therefore, the Paleo-Asian Ocean was likely closed at least in the Early Permian; thereafter large-magnitude strike-slip displacements and related local vertical-axis rotations acted as the major intracontinental tectonics. Moreover, the major blocks in the southern CAOB presented a northwest-southeast distributed paleogeography in the Early Permian.

9.2 Perspectives

Though we have answered important scientific questions during this Ph.D. study, several ones still remain unsolved, calling for further effort to improve our understanding of the tectonic evolution of the CAOB. These are:

1. The southernmost CAOB is a key region to understand the configuration of the supercontinent Pangea and the late Paleozoic to the early Mesozoic tectonic transition from accretion/subduction to intracontinental tectonics. Although our study provides several high-quality paleomagnetic poles for the major blocks in this region, more

paleomagnetic works are needed to reach a deeper comprehension, especially 1) the paleomagnetic data of the South Junggar Block are still poorly evaluated and tested; 2) no robust paleomagnetic data have been reported in the Central Tianshan; and 3) a few late Paleozoic-early Mesozoic paleomagnetic poles were acquired from the Beishan Belt.

2. According to our study, large-magnitude strike-slip displacements and related local vertical-axis rotations acted as the major intracontinental tectonics since or after the Early Permian in the CAO. More detailed field observations, structural analyses, thermochronological studies, and regional comparisons and integrations should be carried out on the fault system in the southernmost CAO. These future works are essential to understand the tectonic evolution of the CAO and will surely shed light on fundamental problems, such as geodynamics of the mega-shear zones and their role on the emplacement of syn-kinematic plutons, strain localization and stress partitioning in regional-scale transtensional and transpressional deformation zones.

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Étude paléomagnétique sur les événements du Paléozoïque supérieur du sud de la Ceinture Orogénique d'Asie Centrale et les contraintes sur ses reconstructions tectonique et paléogéographique

Les évolutions tectonique et paléogéographique du Sud de la Ceinture Orogénique d'Asie Centrale (Central Asia Orogenic Belt-CAOB) du Paléozoïque supérieur au Mésozoïque inférieur présentent une grande importance scientifique mais insuffisamment étudiées jusqu'à maintenant. Les questions clés qui doivent être répondues dans cette thèse pour parvenir à une meilleure compréhension de ces problématiques scientifiques comprennent : identifier la cinématique et de l'évidence géologique de zones en cisaillement en grande échelle; comprendre la transition tectonique de la collision/accrétion à la tectonique intracontinentale; et réaliser une reconstruction paléogéographique du Paléozoïque supérieur. Dans le but d'apporter plus de contraintes, en appliquant la méthode du paléomagnétisme couplée avec d'autres, cette étude a été réalisée sur plusieurs microcontinents du Sud de la COAB. En conséquence, des pôles paléomagnétiques du Permien inférieur de haute qualité ont été obtenus pour les blocs de Yili, Turpan-Hami et Sud de Beishan. En les combinant avec les résultats publiés auparavant, les conclusions suivantes pourraient être tirées: (1) les pôles d'âge Paléozoïque supérieur des principaux blocs de cette région se distribuent le long d'un petit cercle, suggérant que les mouvements relatifs entre ces blocs sont dominés par des rotations relatives, qui pourraient être interprétées par les mouvements importants le long des failles majeures en décrochement; (2) les faibles différences en mouvement latitudinal du Paléozoïque supérieur entre ces blocs principaux confirment et raffinent davantage le modèle tectonique de l'extrusion vers l'est, proposant que la transition tectonique s'est probablement achevée au moins au Permien inférieur; (3) une reconstruction paléogéographique plausible du Permien inférieur du sud de la CAOB a été réalisée, caractérisée par une distribution des blocs majeurs en orientation NW-SE.

Mots clés: Ceinture Orogénique d'Asie Centrale (CAOB), Paléozoïque supérieur, Paléomagnétisme, Reconstruction paléogéographique, Reconstruction tectonique

Paleomagnetic study on the late Paleozoic events of the southern Central Asian Orogenic Belt and constraints on its tectonic and paleogeographic reconstructions

The late Paleozoic to early Mesozoic tectonic and paleogeographic evolutions of the southern Central Asian Orogenic Belt (CAOB) are of great significance but insufficiently investigated. Key issues that need to be resolved for reaching a better understanding of the abovementioned scientific problems contain: revealing the kinematics and geological significance of the large-scale strike-slip shear zones; understanding the tectonic transition from collision/accretion to intracontinental tectonics; and achieving well-constrained late Paleozoic to early Mesozoic paleogeographic reconstruction. Aiming to provide more constraints, applying paleomagnetism coupled with other methods, this study was carried out on several microcontinents in the southern CAOB. As results, high-quality Early Permian paleomagnetic poles for the Yili, Turpan-Hami, and south Beishan blocks were obtained. Combining with published data, the following conclusions could be drawn: (1) the late Paleozoic paleomagnetic poles for the major blocks in this region distribute along a small circle, suggesting that the relative movements among these blocks are dominated by relative rotations, which could be interpreted by the significant strike-slip movements along the major faults; (2) no/weak late Paleozoic latitudinal differences among major blocks confirm and further refine the "eastward extrusion" tectonic model, suggesting that the tectonic transition was probably completed at least in the Early Permian; (3) a plausible Early Permian paleogeographic reconstruction of the southern CAOB was achieved, characterized by a NW-SE distributing pattern of the major blocks.

Keywords: Central Asian Orogenic Belt (CAOB), late Paleozoic, Paleomagnetism, Paleogeographic reconstruction, Tectonic reconstruction



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