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L'évolution tectono-sédimentaire des marges de la Téthys Alpine au cours de l'amincissement lithosphérique

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Douter de tout ou tout croire sont deux solutions également commodes, qui l'une et l'autre nous dispensent de réfléchir.

Henri Poincaré

ABSTRACT

A long-standing problem in Earth Sciences is to understand how continents break apart to form new oceanic basins. Many of the questions that currently frame ongoing debates about continental break-up are related to the mechanics and rheology of extreme lithospheric extension. It is generally accepted that subcontinental mantle is exhumed at magma-poor rifted margins. However, much less attention has been paid to the processes that predate mantle exhumation. An increasing number of observations from magma-poor rifted margins world-wide show evidence for extreme crustal thinning to less than 10km without seismic evidence for significant normal faulting. This leads to the question of what structures/processes can explain such major crustal thinning and where, when and under which conditions these structures/processes are active? The limited resolution of geophysical surveys and the lack of drilling in present-day deep water rifted margins make it difficult to answer to these questions. A more direct access to the sedimentary record of deep water rifted margins and the underlying crust/mantle lithosphere is exposed in the Alps in SE-Switzerland and N-Italy where remnants of the ancient Adriatic rifted margin are preserved. We initiated a research project in the Bernina-Campo-Grosina units in SE-Switzerland and N-Italy, which exposes remnants of the transition between the proximal and the distal Adriatic rifted margin. The major results obtained from the investigation of these units are:

1. The characterization of the Alpine structures and the reconstruction of the Jurassic rift architecture in the Bernina-Campo-Grosina units enable to show that the Bernina unit is preserving relics of a former distal margin while the Campo-Grosina units form relics of a former necking zone.
2. The necking zone is characterized by exhumation of middle crustal levels accommodated by a system of conjugate crustal scale detachment systems that interact with middle crustal decollements resulting in the thinning and omission of middle crust.
3. In the distal margin, upper crustal rocks are juxtaposed against lower crustal and mantle rocks along extraction faults during extreme crustal thinning of the future distal margin. Final mantle exhumation is associated with detachment faults that cut across the residual, hyper-extended brittle crust resulting in exhumed detachment faults that are overlain by extensional allochthons and sealed by syn-rift sediments.

These new results permit to understand the extreme thinning of the crust during final rifting as well as to propose a coherent rift evolution of the Alpine Tethys margin and last but not least, to unravel the importance of rift inheritance for the subsequent Alpine reactivation.

RÉSUMÉ

La théorie de la tectonique des plaques implique des processus d'extension de la lithosphère amenant à la déchirure continentale puis à la création d'un domaine océanique couplé avec une nouvelle limite de plaque. Cependant l'ensemble des mécanismes d'extension de la lithosphère qui vont engendrer cet amincissement reste encore mal compris. Depuis les années 1960-1970 et les premiers modèles conceptuels visant à comprendre l'extension lithosphérique, de nombreuses découvertes ont été effectuées dans les marges passives peu magmatiques révélant des géométries complexes. La majorité des travaux s'est focalisée d'une part sur les parties proximales des marges ayant une lithosphère peu ou pas amincie ou d'autre part sur les Transitions Océan-Continent (TOC). Les résultats sur les TOC ont démontré l'exhumation de manteau subcontinental ainsi que de croûte continentale grâce à l'action de faille de détachement à faible pendage lors de la fin du rifting. L'acquisition de nombreuses nouvelles données sur les marges passives peu magmatiques (Norvégienne, Sud Atlantique, Ibérique) ont mis en évidence que, lors du jeu des failles de détachement à faible pendage, la croûte a déjà subi un amincissement majeur et ne possède plus qu'une épaisseur voisine de 10 km, parfois moins, sur des distances qui peuvent atteindre jusqu'à 200 km. Ceci implique que ces failles n'ont accommodé que la toute dernière phase d'amincissement qui reduira l'épaisseur de la croûte de 10 km à 0, avant la déchirure continentale. Les structures imagées sismiquement et les données de forages lors de campagne ODP ne montrent pas des structures d'extension capables d'accommoder cet amincissement, et dans certains cas l'extension se produit sans aucune structure visible en coupe sismique. Les observations géophysiques sur les parties proximales des marges ne montrent en revanche qu'un amincissement tout relatif avec une croûte peu (de l'ordre de 1 à 2 km) ou pas amincie.

Ainsi, les processus qui vont accommoder l'amincissement crustal de 30 km à 10 km restent pour l'instant méconnus. Le but de ma thèse est de comprendre les mécanismes qui vont induire le passage à une croûte relativement peu amincie dans la marge proximale à une zone extrêmement amincie dans les parties distales des marges. Quels sont les processus et les mécanismes qui vont accommoder cet amincissement et cette extension lithosphérique ? Quelle est la réponse du système sédimentaire ? Quelle est la durée de cette déformation ?

Les réponses à ces questions ne peuvent pas être trouvées dans les marges distales actuelles qui sont de fait déjà amincies. Cependant, cet extrême amincissement est enregistré dans les zones d'étranglement ou « necking zone ». Dans ces zones, en coupe sismique, il est possible d'observer la remontée brutale du Moho sur une distance de 50 à 60 km avec une

épaisseur crustale qui passe de 30 km à 10 km. La difficulté d'identifier avec précision l'architecture crustale ainsi que les structures d'extension à l'aide de sismique réfraction et réflexion, couplée avec l'absence de forage ODP dans les « necking zones », rendent les marges passives actuelles peu favorables à la résolution de ces questions. En effet pour comprendre l'amincissement, il est nécessaire d'avoir accès à l'architecture crustale ainsi que de pouvoir contraindre les structures d'extension.

Ainsi, ma thèse concerne plus particulièrement les marges fossiles Téthysiennes Alpines. Ce domaine bénéficie de plus de 100 ans de travaux géologiques et constitue un laboratoire naturel unique pour étudier les marges passives peu magmatiques. Grâce aux précédents travaux, la position paléogéographique des différentes unités est contrainte, permettant l'identification des différentes parties de la marge. Néanmoins, aucune « necking zone » n'avait été identifiée. Dans le but de comprendre les mécanismes d'extension lithosphérique, mon étude c'est localisée dans la marge fossile Adriatique, affleurant actuellement dans les nappes Australpines et Sud Pénninique dans le Sud-Est de la Suisse et le Nord de l'Italie. Ces nappes préservent les reliques complètes d'une ancienne marge passive peu magmatique. Les études précédentes sur la zone ont caractérisé l'enchaînement des différentes phases de déformation Alpines. Trois phases majeures ont été identifiées mais seul le premier événement tectonique du Crétacé supérieur a eu une incidence majeure. Il a provoqué l'empilement des nappes suivantes, du continent vers l'océan et du plus haut structuralement au plus bas: (1) les nappes Austroalpines supérieures représentant les reliques d'une marge proximales (2) les unités de Bernina-Campo/Grosina et (3) les nappes Austroalpines inférieures et Sud Pénnique provenant respectivement de la marge distale et de la TOC. Les unités de Bernina-Campo/Grosina n'ont quand à elles fait l'objet d'aucune étude précise et leurs positions paléogéographiques n'ont été assignées à aucune partie de la marge. Ces unités occupent une position remarquable, car elles se situent entre des reliques de marge proximale et distale, c'est-à-dire à la possible position de la « necking zone ».

L'étude comporte deux chantiers distincts apportant chacun un niveau de compréhension différent pour les analyses des mécanismes d'extension lithosphérique. La première partie de ma thèse se concentre dans la marge fossile Adriatique. Dans ce chantier, je traite de la géologie Alpine de ma zone d'étude puis des structures issues du rifting Jurassique et de leurs implications dans l'amincissement crustal. Dans une seconde partie, l'étude est élargie au domaine Téthysien Alpin dans sa globalité, avec pour but de proposer un modèle conceptuel de l'extension lithosphérique des marges passives peu magmatiques. Ainsi, ce travail est articulé autour de deux axes de résultats majeurs : (1) La compréhension et la description de la déformation Alpine, pré-requis nécessaire en vue d'accéder à une reconstruction palinspastique cohérente, (2) la reconnaissance et la caractérisation de la géométrie pré-Alpine et plus particulièrement celle issue du rifting, en vue d'expliquer les mécanismes d'amincissement lithosphérique.

Réactivation et évolution Alpine de la marge Adriatique, nouvelles observations

La première partie de ma thèse s'est focalisée sur la géologie Alpine des nappes Austroalpines et Sud Pénnique dans le Sud-Est de la Suisse et le Nord de l'Italie. Grâce à un travail de cartographie détaillé couplé aux précédentes études, une nouvelle caractérisation des différents événements tectoniques a été proposée. Il est présenté dans cette étude une nouvelle interprétation concernant l'imbrication des nappes Austroalpines. La zone d'étude a été affectée par une succession de phase de déformation en relation avec deux systèmes compressifs distincts, l'un au Crétacé supérieur et l'autre au Tertiaire. Les nouveaux travaux effectués ont permis de montrer le caractère primordial de la pré-structuration issue du rifting pour la réactivation en compression de la marge. En effet, les structures du rift vont être réutilisées induisant une réactivation hétérogène pendant la déformation Alpine. Ceci va avoir pour conséquence la préservation de certaines domaines et la réactivation d'autres (e.g les unités de Margna-Sella-Malenco). Deux zones de réactivation majeures ont été identifiées (zones d'Albula-Zebru et de Lunghin-Mortirolio) représentant des structures préexistantes issues de l'extension lithosphérique. Les reconstructions mettent en évidence le rôle important de l'héritage structural sur les processus de réactivation.

Les nouvelles observations effectuées plus particulièrement sur les unités de Bernina-Campo/Grosina ont permis de proposer des reconstructions palinspastiques et paléogéographies cohérentes et d'assigner une position paléogéographique précise pour ces unités. Dans l'unité de Bernina ont été reconnues des structures et une géométrie d'extension caractéristique de la marge distale. Les unités de Campo/Grosina sont interprétées comme occupant les reliques d'une ancienne « necking zone ».

Caractérisation de l'architecture crustale et sédimentaire de la marge distale

L'unité de Bernina représentant les reliques d'une marge distale, est essentiellement composée de socle de croûte supérieure pré-rift avec quelques écailles de sédiments mésozoïques (pré- et syn-rift). Une cartographie détaillée de cette zone établit la présence au toit du socle d'une zone de faille composée de cataclasites et de gouges. Celle-ci montre un faible pendage et est localement surmontée par des fragments de dolomites pré-rift d'âge Triasique. Ceux-ci sont surmontés à plat par des calciturbidites syn-rift du Jurassique moyen qui scellent le socle, montrant localement des onlaps sur les sédiments triasiques. Cette relation permet de démontrer la présence d'une faille de détachement à faible pendage qui accommode l'exhumation de croûte supérieure à la fin du rifting. Cette faille de détachement peut être cartée dans toute la zone au toit du socle, et peut être suivie localement à la limite entre les sédiments du Permo-Trias et le socle paléozoique. Ces relations mettent en évidence la présence d'allochtone extensionnel de pré-rift ainsi que l'exhumation de croûte supérieure pendant le rifting. Ces nouvelles observations traduisent l'accommodation de l'extension dans les marges distales par des failles de détachement à faible pendage provoquant la délamination des sédiments pré-rift et l'exhumation de croûte supérieure et de manteau subcontinentale exhumé dans les TOC.

Les études antérieures sur l'unité de Margna peuvent nous apporter des contraintes quand à l'architecture crustale de la croûte dans la marge distale. En effet, dans cette unité a été décrite la juxtaposition d'une croûte supérieure et d'une croûte inférieure le long de structures cisaillantes majeures, actives lors du rifting. Les études pétrologiques menées sur cette zone ont démontré qu'il manquait jusqu'à 20 km de croûte continentale, essentiellement de la croûte moyenne. Ces observations permettent de montrer que la marge distale dans la zone étudiée est principalement composée de croûte supérieure et inférieure avec l'omission de croûte moyenne. Ainsi la marge distale semble composer des niveaux les plus supérieurs et inférieurs de la croûte. Ces déformations sont localisées le long de zone de cisaillement dans la croûte inférieure ou le long de détachement à faible pendage dans la croûte supérieure.

Nouvelles données pour la caractérisation des processus d'extension lithosphérique : le rôle majeur de la necking zone

Les reconstructions qui ont été effectué montrent que les unités de Campo et Grosina représentent une ancienne « necking zone ». La déformation Alpine est très faible et localisée à quelques horizons tectoniques. L'unité Campo se compose majoritairement de métapélites équilibrées dans le faciès Amphibolite supérieur, montrant une foliation E-W remarquablement constante dans la zone étudiée. Cette foliation est coupée par l'intrusion du gabbro Permien de Sondalo. Des études antérieures ont permis de montrer que ce gabbro s'est mis en place entre 270 et 300 Ma à une pression de 0,8 Gpa et à des températures voisines de 900°C. Ainsi l'unité de Campo représente une portion de croûte moyenne pendant le Permien. Du Permien au Trias, aucune trace d'activité tectonique n'a été observée. Il est donc raisonnable de penser qu'au début du Jurassique ces roches résidaient toujours dans la croûte moyenne. Cette configuration permet la caractérisation de son évolution pendant le rifting.

Dans ce but, des études thermochronologiques avec la méthode $^{40}Ar/^{39}Ar$ ont été entreprises sur les roches encaissantes de l'unité de Campo au voisinage de l'intrusion de Sondalo. Les biotites fournissent des âges de refroidissement de 184.55±1.22 Ma à 188.67±0.86 Ma indiquant que ces roches ont été exhumées à des températures inférieures à 350-320°C à la fin du rifting. Ces résultats indiquent que la «necking zone» est caractérisée par l'exhumation de roche profonde pouvant provenir de la croûte moyenne.

Les structures d'extension accommodant cette exhumation ont été révélées au toit de l'unité de Campo. Une zone de cisaillement majeure à faible pendage appelée la zone de cisaillement d'Eita, marque la transition entre l'unité de Campo représentant la croûte moyenne et l'unité de Grosina relique d'une croûte supérieure. Cette zone avait été interprétée antérieurement comme étant liée à une déformation Alpine. Cependant, les nouvelles études cartographiques, structurales, métamorphiques ainsi que thermochronologiques ne sont pas en accord avec une origine Alpine de cette structure. En me basant sur ces nouvelles données, je propose que cette zone de cisaillement soit liée au mécanisme d'exhumation de la croûte moyenne pendant le rifting. Cette zone de cisaillement est caractérisée par une déformation comprise entre 300 et 400°C. La base de l'unité de Grosina qui se situe au contact de la zone de cisaillement est marquée par une déformation pénétrative et rétrograde sur plus de 200 m dans le faciès « schiste vert » et est associé à une foliation plate. Cette foliation, étant en relation avec la zone de cisaillement d'Eita, est interprétée comme permettant l'extrusion latérale de la croûte moyenne.

Plus haut dans l'unité de Grosina, on retrouve une foliation pré-Alpine similaire à celle enregistrée dans l'unité de Campo. Le toit de l'unité de Grosina est caractérisé par la présence d'une zone de déformation cassante à géométrie plate (<20°) assimilée à une faille de détachement (détachement de Grosina). L'absence de déformation plastique et de recristallisation dynamique du quartz permet de montrer que cette faille a été active à des températures inférieures à 300°C. Néanmoins, le fait que cette faille soit intra-socle et qu'il n'y ait pas de trace de sédiments mésozoïques rend difficile la détermination de l'âge exact de cette structure. Cependant, localement l'unité de Campo est juxtaposée à l'unité de Bernina le long d'un contact cassant précoce qui semble être réactivé et recoupé par la première phase de déformation Alpine. Ce contact est interprété comme étant la prolongation du détachement de Grosina qui favorise l'exhumation de croûte moyenne pendant le rifting, aidé par la zone de cisaillement d'Eita qui, quand à elle, accommode l'extrusion latérale. Ces nouvelles observations ont permis l'identification et la description des structures accommodant l'extension lithosphérique pendant le rifting à la fin du Lias et au début du Dogger (~180 Ma). Ces observations permettent ainsi de caractériser les processus d'amincissement actif dans la «necking zone». Les unités de Campo-Grosina offrent la possibilité de faire une coupe à travers une ancienne «necking zone» ayant accommodé l'exhumation et l'extrusion de croûte moyenne pendant le rifting.

Le domaine Téthysien Alpin : un laboratoire naturel pour comprendre l'extension lithosphérique

La dernière partie de mon étude traite de l'ensemble du système Alpin, en vue de le reconstruire. Les phases précoces du rift, affectant principalement les marges proximales Adriatique et Européenne du Trias (~200 Ma) jusqu'au Pleinsbachien-Toarcien (~ 180 Ma) sont bien contraintes. Il va de même pour l'évolution tardive du rift avec l'exhumation de manteau subcontinental. En revanche, l'amincissement crustal et la géométrie des parties distales des marges restent méconnus. Grâce aux nombreux travaux effectués dans les Alpes, il est possible de corréler les données stratigraphiques ainsi que les données structurales et pétrologiques. Ces études antérieures couplées aux nouvelles données de cette thèse permettent de proposer un modèle cohérent pour l'évolution des différentes parties du domaine Téthysien pendant le rifting. Pendant la phase d'amincissement, les necking zones Européenne et Adriatique vont individualiser le domaine Briançonnais. Ce domaine va subir une évolution atypique. En effet, alors que les domaines adjacents sont marqués par une subsidence importante, le domaine Briançonnais va être soulevé et rester en milieu sub-aérien. A la fin du rifting, le bloc Briançonnais va être délaminé par des failles de détachement à faible pendage, qui vont induire la création des marges distales de par et d'autre du bloc. Les «deux necking zones» des deux marges jouent le rôle primordial de découpleur, qui va permettre l'individualisation du bloc Briançonnais et sa délamination. Ces mécanismes vont permettre la création des marges distales avec une croûte continentale extrêmement amincie. Ceci implique qu'antérieurement au rifting, les domaines distaux des marges Européenne et Adriatique n'existaient pas et ne sont que le produit de la délamination du bloc Briançonnais lors de l'extension

Les résultats de ce travail ont permis de contraindre l'architecture crustale de la «necking zone» ainsi que celle des parties distales des marges. Ce travail a permis pour la première fois la reconnaissance et la caractérisation d'une necking zone qui n'avait été jusqu'alors uniquement imagée sur coupes sismiques. En effet, les reliques de cette zone, exposée dans les unités de Campo-Grosina, permettent de caractériser l'exhumation et l'évolution d'un niveau crustal intermédiaire pendant le rifting. Le comportement de la croûte moyenne n'avait été que très peu décrit car mal identifié, la majorité des études se focalisant sur les niveaux supérieurs ou inférieurs. En revanche, les marges distales sont caractérisées par la juxtaposition de croûte supérieure et inférieure avec la disparition de la croûte moyenne. Cette étude a permis la description des structures responsable de l'extrême amincissement crustal à la fin du rifting. Les observations ont permis de proposer un modèle global capable d'expliquer les processus d'amincissement amenant à la déchirure continentale. Ces résultats ont des implications majeures pour la compréhension des processus d'extension lithosphérique dans les marges actuelles amenant à l'océanisation. De plus, ces travaux montrent le rôle essentiel de l'héritage structural et de la pré-structuration issus du rifting pour la compréhension de la réactivation et de l'imbrication des marges dans un système orogénique.

En conclusion, dans l'évolution des Alpes, une étroite relation existe entre le système géodynamique de rift Jurassique et les systèmes Crétacé et Tertiaire liés aux fermetures des domaines océaniques. Ces deux événements tectoniques sont couplés. La compréhension des orogènes actuelles nécessite la connaissance de la complexité structurale préexistante issue du rifting. En revanche, la description du rifting implique l'analyse et le décryptage des processus de compression. La comparaison et la confrontation des analyses, effectuées conjointement sur les marges passives et sur les orogènes, semblent être une clé pour une meilleure compréhension de ces deux phénomènes.

REASSUNT

Ün problem da lunga dürada in scienzas da la terra es l'incletta sco cha continents as separan et oceans as fuorman. Las discussiuns chi fan part dal stüdi sur la rottadüra e la separaziun da continents sun liats als mechanisms d'extensiun extrema da la cruosta continentala als urs dals continents. Las observaziuns als urs dals continents actuals muossan cha'l passagi d'üna cruosta da grosseza normala a une cruosta fina passa par una zona da asutiglimaint (ausdünnen). Il stüdi da quista zone es pero fich dificila, a causa dal fat cha'ls urs dals continents as rechattan sout il fond dal mar, inaccessibel à l'observaziun directa. Per stüdiar ils process chi sun associats à la separaziun dals continents, vaina stüdià las unitats da Bernina – Campo – Grosina, chi pertoccan a las cuvertas da las Alps occidentalas. In quistas cuvertas as rechattan vegls relicts dal ur dal continent adriatic. Nos stüdi a permis da descriver rotaduras liadas a l'asutiglimaint (ausdünnen) dal la litosfera continentala chi's preschaintan in fuorma da : 1) rottadüras plattas activas aint illas parts resistentas da la cruosta, 2) zones ductilas tanter las parts resistentas da las cruosta superiuras et inferiuras, e 3) zones d'extracziun responsablas pel asutiglimaint da las parts ductilas da la cruosta. L'interacziun das quists process han permis da asutigliar (ausdünnen) la cruosta da 30 km a main da 10 km par l'extracziun da material ductil durant il Toarcian (180 milliuns d'ons) al fond dal mar alpin. Plü tard, vers 165 milliuns d'ons, cur cha la cruosta d'eir fingia finna (< 10 km), rottaduras plattas han pudü traversar la cruosta e penetrar aint il mantel suot la cruosta e til tirar al fond dal mar. Quistas observaziuns fattas aint illas Alps permettan da reconstruir l'architectura dal urs dals continents, da studiar ils process liats a lur fuormaziun e da congualar (vergleichen) ils resultats cun observaziuns indirectas fattas par metodas da geophysica da quistas zonas chi sun actualmaing a millis da meters sout l'ova, tanter ils continents e'ls oceans inaccessibel a l'observaziun directa da l'uman.

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INTRODUCTION

1. L'évolution des concepts et des modèles structuraux sur les marges passives peu magmatiques, des années 1960 à nos jours

Depuis l'établissement de la théorie de la tectonique des plaques, les évolutions et changements de concepts et d'hypothèses sur les processus de rifting et de drifting ont été nombreux. Ces évolutions sont le fruit d'une étroite interaction et confrontation entre les résultats des travaux effectués en mer sur les marges passives actuelles, principalement dans le système Ibérie-Terre Neuve et les études réalisées à terre sur les marges fossiles de la Téthys Alpine (Fig. 1). La connaissance et la compréhension des processus conduisant à la rupture continentale ainsi que de ceux responsables de l'extension lithosphérique restent à l'heure actuelle un des enjeux fondamental des sciences de la Terre. Ces questions sont situées à l'interface de nombreuses disciplines telles que : géologie de terrain, géologie, géochimie, géophysique marine et modélisation numérique.

Dans la suite de ce chapitre, j'exposerai les avancées des connaissances sur les marges passives peu magmatiques ainsi que les différentes évolutions des concepts, pour ensuite introduire le sujet qui est traité dans ce mémoire de thèse.

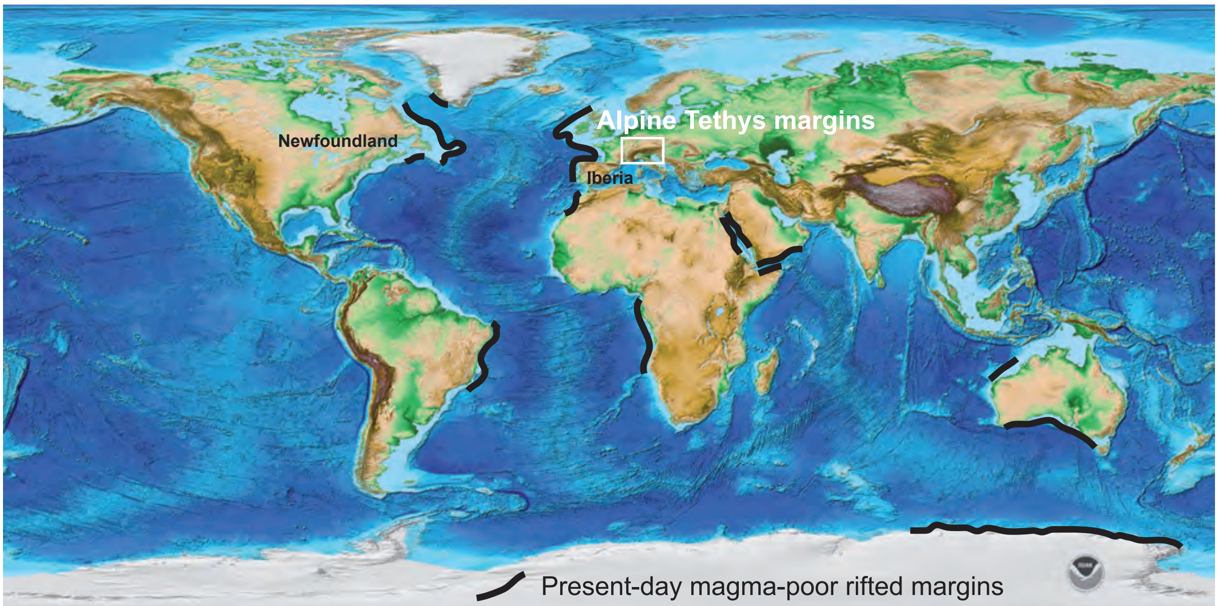


Fig. I. 1: Modèle numérique de terrain montrant les positions des reliques des marges de la Téthys Alpine et des marges passives actuelles, en particulier le système Ibérie-Terre Neuve. Image modifiée de NOAA, National Geophysical Data Center.

1. 1. 1960-1980 : Vers un modèle général pour la compréhension du rifting continental

Les nombreux travaux effectués sur la chaîne Alpine (Fig. 1) ont conduit les chercheurs à démontrer que l'établissement de l'orogène s'est effectué au dépens de bassins sédimentaires préexistants (e.g. Bertrand 1894 ; Argand 1916 ; Staub 1917 ; Bernoulli 1964 ; Trümpy 1984 ; Lemoine et al. 1986 ; Schmid et al. 2004 et références citées), même si la transposition de ces phénomènes à un schéma géodynamique global resta longtemps mal comprise. À l'inverse, les mécanismes pouvant accommoder la création de ces bassins sédimentaires ont très vite été identifiés. Ainsi de nombreux travaux (e.g. Günzler-Seiffert 1942 ; Bernoulli 1964 ; Trümpy 1984) ont pu décrire et comprendre la formation de bassins contrôlée par le jeu de failles normales à fort pendage (Fig. 2). Ces études ont été complétées par les progrès de la géologie marine qui ont permis dès les années 1970 d'imager le premier bloc basculé de la marge Armoricaine actuelle (Fig. 2) (De Charpal et al. 1978 ; Montadert et al. 1979). Ces découvertes ont été très vite suivies par l'analyse de structures similaires dans le domaine Téthysien Alpin (Fig. 2) (Bernoulli 1964, 1979 ; Lemoine et al. 1986 ; Lombardo et Pognante 1982). Ainsi, la comparaison des données de terrain et celles acquises en mer ont permis une nouvelle avancée de la compréhension de l'architecture et de l'évolution des bassins sédimentaires, de type bloc

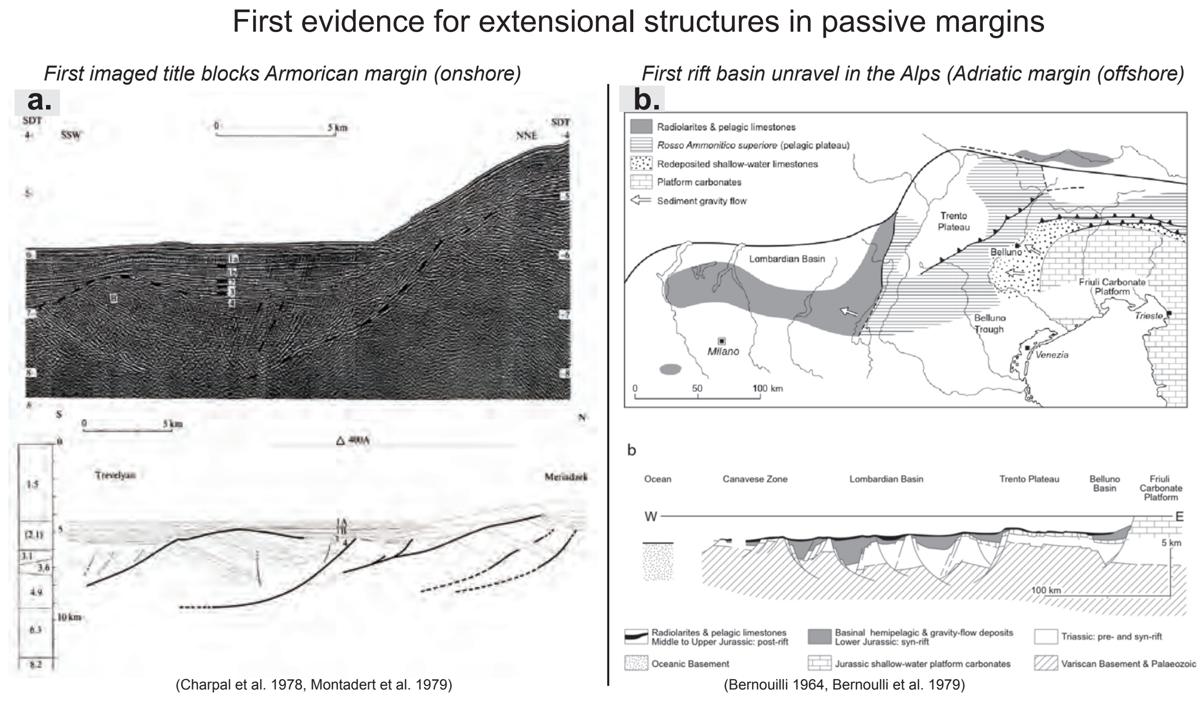
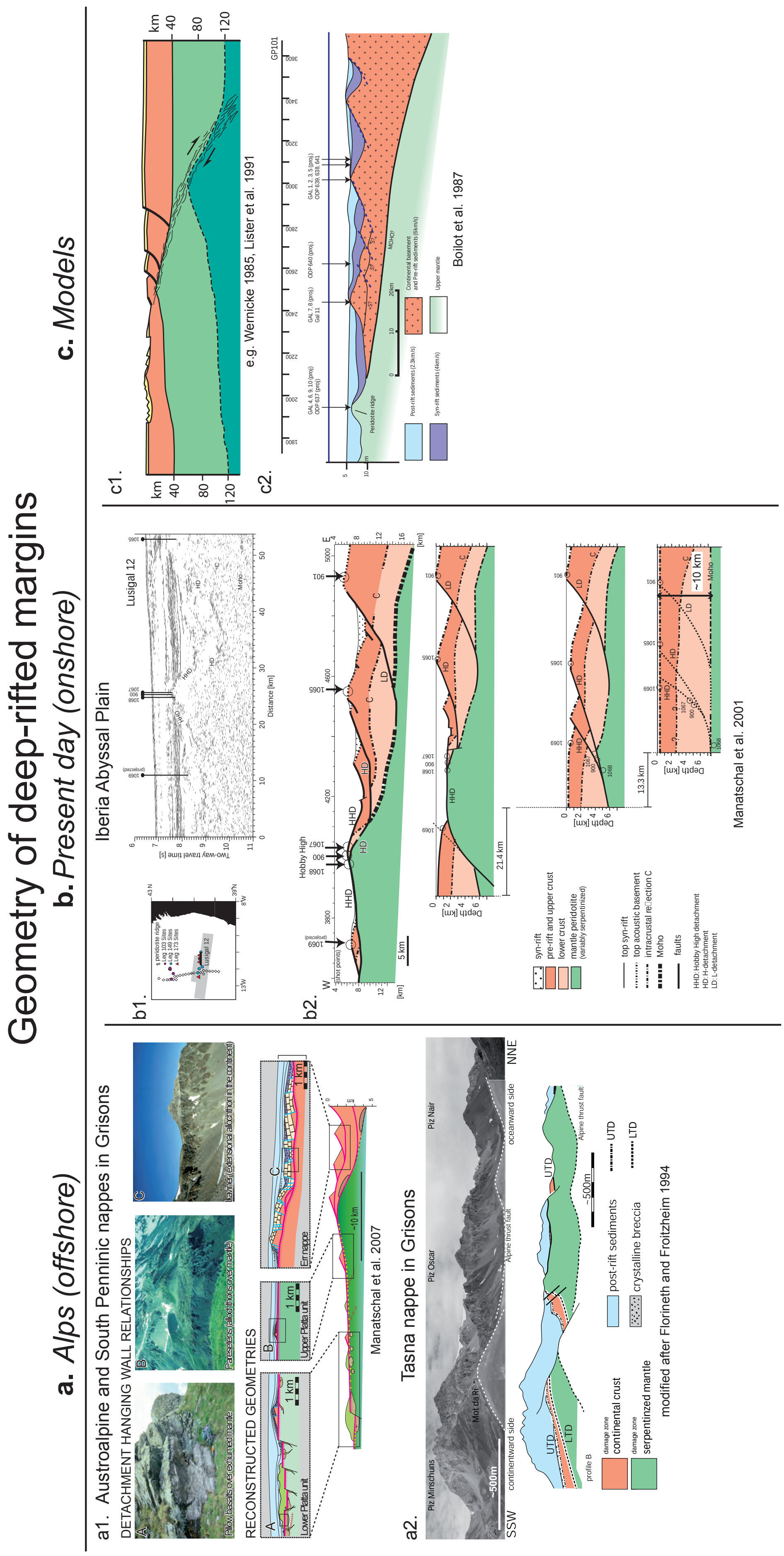


Fig. I. 2: Figure montrant les premières structures extensives en liaison avec le rifting reconnu (a) dans la marge passive Armoricaine sur sismique réflexion (De Charpal et al. 1978 ; Montadert et al. 1979) ; (b) dans les Alpes dans les reliques de la marge passive Adriatique. Une cartographie détaillée de l'architecture des dépôts Mésozoïques a permis d'identifier les bassins de rift (Bernoulli 1964 ; Bernoulli et al. 1979).

basculé, contrôlés par des failles normales à fort pendage. C'est dans ce contexte de recherches que Mc Kenzie en 1978 proposa un modèle cohérent de l'extension lithosphérique expliquant l'amincissement crustal qui mène à l'initiation d'un centre d'accrétion océanique. Ce modèle est basé principalement sur le fait que l'amincissement est uniforme et constant en fonction de la profondeur (βcroûte=βManteau). L'extension de la lithosphère se produit suite à un régime tectonique en cisaillement pur, et a pour implication une relative symétrie des deux marges conjuguées. Ce type de modèle a été utilisé avec succès principalement pour l'étude des bassins formés par des blocs basculés et typiquement localisés dans les marges proximales et dans de nombreuses marges actuelles peu magmatiques.

1. 2. 1980-2000: L'extension lithosphérique est fonction de la profondeur (Depth-Depend-Stretching, DDS) : nouvelles observations et modèles

Au milieu des années 1980 apparut le modèle de Wernicke (1981, 1985) développé à partir de la structure des « métamorphic core complexes » de la province du « Basin and Range » aux Etats-Unis. Il propose la présence de détachement à faible pendage d'échelle lithosphérique pour accommoder l'extension pendant le rifting. Ce modèle implique une extension contrôlée par des processus en cisaillement simple qui provoque l'asymétrie des deux marges conjuguées. À la même époque la présence de manteau subcontinental sous la couverture sédimentaire post-rift est découverte au large de la Galice dans la marge Ibérique (Fig. 3) (Boilot et al, 1987). Cette nouvelle observation a eu des conséquences majeures sur le développement des nouveaux concepts d'extension lithosphérique et de déchirure continentale. Elle va être confortée et développée grâce aux travaux effectués à terre dans les marges de la Téthys Alpine où des structures semblables ont été identifiées (Fig. 3) (Froitzheim et Eberli 1990, Manatschal et Nievergelt 1997). Ces études menées conjointement à terre et en mer ont permis de mettre en évidence la présence de faille de détachement, à faible pendage, situées au toit de la croûte continentale dans les parties distales, et qui coiffent le manteau subcontinental exhumé dans les Transitions Océan-Continent (TOC). Ainsi le modèle de faille de détachement à faible pendage d'échelle lithosphérique (Wernicke 1981, 1985 ; Lister et al. 1986 ; Lister et Davis 1989) a pu être appliqué à l'exhumation de manteau subcontinental dans les TOC de la marge Ibérique actuelle (Fig. 3) (e.g. Manatschal et al. 2001) ainsi qu'aux marges fossiles de la Téthys Alpine (Lemoine et al. 1987). En outre, le mécanisme d'exhumation proposé permet d'expliquer l'évolution et la structure asymétrique caractérisant les deux marges conjuguées à la fin du rifting, aussi bien celles des marges actuelles que celles des marges fossiles. Cependant, la caractérisation des failles de détachement accessibles dans la chaîne Alpine a permis de démontrer qu'elles sont actives uniquement dans le domaine cassant et qu'elles ne montrent aucune relation de conti-


Geometry of deep-rifted margins
a. Alps (offshore)
b. Present day (onshore)
c. Models
a1. Austroalpine and South Penninic nappes in Grisons
DETACHMENT HANGING WALL RELATIONSHIPS
RECONSTRUCTED GEOMETRIES
Lower Platta unit
Upper Platta unit
Err nappe
1 km
~10 km
Manatschal et al. 2007
a2.
Tasna nappe in Grisons
Piz Minschuns
Mot da Ri
Piz Oscar
Piz Nair
SSW
NNE
continentward side
oceanward side
Alpine thrust fault
UTD
LTD
~500m
post-rift sediments
crystalline breccia
damage zone
continental crust
serpentinzed mantle
modified after Florineth and Froitzheim 1994
b1.
Iberia Abyssal Plain
Lusigal 12
Two-way travel time [s]
Distance [km]
b2.
Hobby High
HHD
HD
LD
5 km
Depth [km]
21.4 km
13.3 km
~10 km
syn-rift
pre-rift and upper crust
lower crust
mantle peridotite (variably serpentinized)
top syn-rift
top acoustic basement
Moho
faults
HHD: Hobby High detachment
HD: H-detachment
LD: L-detachment
Manatschal et al. 2001
c1.
km
40
80
120
e.g. Wernicke 1985, Lister et al. 1991
c2.
GP101
Peridotite ridge
Continental basement and Pre-rift sediments
Upper mantle
Post-rift sediments
Syn-rift sediments
Boillot et al. 1987

nuité avec une déformation mylonitique (Manatschal 2004). Ces faits indiquent en opposition aux données sur les « metamorphic core complexes » et aux modèles expliquant l'extension lithosphérique (Wernicke 1981, 1985 ; Lister et al. 1986 ; Lister et Davis 1989), que dans un système de rift, ces structures sont restées actives dans des conditions de températures inférieures à 300°C. Ces conclusions ont été confirmées par les campagnes de forage ODP Leg 103, 147, 173 et 210 sur la marge Ibérique (Whitmarsh et al. 1996). En effet, les données de forages mises en corrélation avec celles des coupes sismiques effectuées sur la marge distale et dans la TOC ont permis de caractériser les mécanismes de l'extension de ces zones. Les marges proximales montrent une géométrie contrôlée par des failles à fort pendage avec une exhumation limitée au voisinage de l'escarpement de faille, tandis que les marges distales montrent une extension contrôlée par des failles de détachement à faible pendage en relation avec une exhumation de croûte continentale et une exhumation de manteau dans les TOC (Manatschal 2004) (Figs. 2, 3). Dès lors, la plupart des études se sont focalisées sur les domaines distaux avec l'objectif de caractériser avec précision les processus actifs dans ces zones, tels que : (1) les mécanismes de l'extension permettant l'exhumation du manteau (e.g. Whitemarsh et al. 2001 ; Manatschal et al. 2007) (2) le magmatisme dans les TOC (e.g. Minshull et al. 2001 ; Müntener et Manatschal 1996) (3) la structure géophysique des TOC dans les marges actuelles, (e.g. Whitmarsh et al. 2001) (4) et l'importance de la serpentinisation (e.g. Pérez-Gussinyé et al. 2001). Les données de ces travaux tendent à montrer que contrairement au modèle établi de rift (modèle type de Mc Kenzie 1978), la transition entre la croûte continentale et la croûte océanique n'est pas abrupte mais plutôt progressive avec un domaine intermédiaire contenant du manteau subcontinental exhumé pouvant atteindre jusqu'à 170 km de longueur (Whitmarsh et al. 2001). De plus, les structures accommodant l'extension entre les parties proximales et distales des marges sont très différentes. Ainsi, ces nouveaux modèles structuraux, construits aussi bien pour les marges actuelles que pour les marges fossiles, montrent que les précédents modèles de Mc Kenzie (1978) et Wernicke (1985) ne peuvent plus rendre compte de tous les faits d'observations relevés dans les marges passives peu magmatiques et sont par conséquent incomplets.

Fig. I. 3: (a1) Reconstruction palinspastique de la Transition Océan-Continent de Err-Platta, montrant les relations structurales entre le mur et le toit des détachements à faible pendage accommodant l'extension lors du rifting Jurassique (figure d'après Manatschal et al. 2007). (a2) Panorama de la Transition Océan-Continent dans la nappe Tasna. Il est possible sur cet affleurement de visualiser le passage d'une croûte amincie jusqu'au manteau exhumé sans réactivation Alpine (figure d'après Florineth et Froitzheim 1994). (b1) Carte de la marge passive Ibérique et localisation de la coupe sismique Lusigal 12. (b2) Evolution spatiale et temporelle de la marge distale Ibérique. Il est important de noter que si l'on rétrodéforme les structures extensives visibles, la croûte continentale possède déjà une épaisseur de 10 km (Manatschal et al. 2001).

1. 3. 2000: L'amincissement crustal, un enjeu majeur pour la compréhension de l'évolution des marges passives peu magmatiques

La fin des années 1990 et le début des années 2000 marquent le passage de travaux focalisés sur les TOC vers les domaines distaux des marges. Ils ont pour but de comprendre la géométrie de ces domaines ainsi que les processus qui vont initier l'amincissement crustal, puis la rupture continentale. Les travaux de Manatschal et al. (2001) et Péréz Guissnyé et Reston (2001) établissent que dans les parties distales de la marge Ibérique, la rétro-déformation des structures d'extension, aussi bien à faible que à fort pendage et imagées sur les coupes sismiques ainsi que par forages, définit une croûte continentale déjà amincie à une épaisseur voisine de 10 km. Ceci a une implication majeure : l'exhumation de manteau subcontinental le long de faille de détachement ne représente qu'un événement tardif dans l'histoire du rift (Manatschal 2004, Lavier et Manatschal 2006). Cette conclusion implique que la croûte continentale a dû être amincie lors d'un événement antérieur. Ainsi lorsque les failles de détachement sont actives, la croûte se trouve déjà entièrement dans le domaine cassant. Ce mécanisme permettra le couplage entre croûte continentale et manteau subcontinental, le long d'un détachement actif lors d'une phase tardive du rifting (Manatschal 2004 ; Reston 2009). Des études récentes ont soulevé de nombreuses questions concernant les parties distales des marges :

(1) dans les parties distales, l'extension établie par les structures imagées par la géophysique implique un amincissement plus faible que celui observé (Fig. 4) (e.g. Contrucci et al. 2004 ; Moulin et al. 2005 ; Kusznir et Karner 2007, Reston 2009); de plus dans certains cas la croûte a été amincie sans indice visible de faille normale (Fig. 4) (e.g. Sibuet 1992, Driscoll et Karner 1998),

(2) la subsidence observée est plus faible que celle attendue dans un système où la croûte a subi un amincissement extrême (Kusznir et Karner 2007),

(3) à la fin du rifting, les marges conjuguées semblent marquées par une brutale asymétrie tandis qu'au début du rift elles sont plutôt marquées par une symétrie (Reston 2009),

(4) l'architecture crustale de la croûte dans les zones distales et le changement rhéologique durant l'amincissement restent à ce jour mal établis (Lavier et Manatschal 2006).

2. Les questions et objectifs scientifiques de cette étude

L'évolution des connaissances et des concepts sur les marges passives peu magmatiques établie par les résultats des travaux effectués en mer et sur terre a pointé des questions restant à résoudre, en particulier celles sur les mécanismes de l'extension lithosphérique dans les parties distales des marges.

En effet, les études menées dans les années 70 ont considérablement contribué à la connaissance des géométries des parties proximales. Celles-ci sont contrôlées par des failles normales créant une géométrie classique en blocs basculés mais n'induisant pas un amincissement conséquent de la croûte. Les études récentes de sismique réfraction dans la marge de Terre Neuve ont démontré que la croûte reste relativement peu amincie dans cette zone. En effet, sur la marge de Terre Neuve, la croûte reste à une épaisseur de 30 à 40 km dans les parties

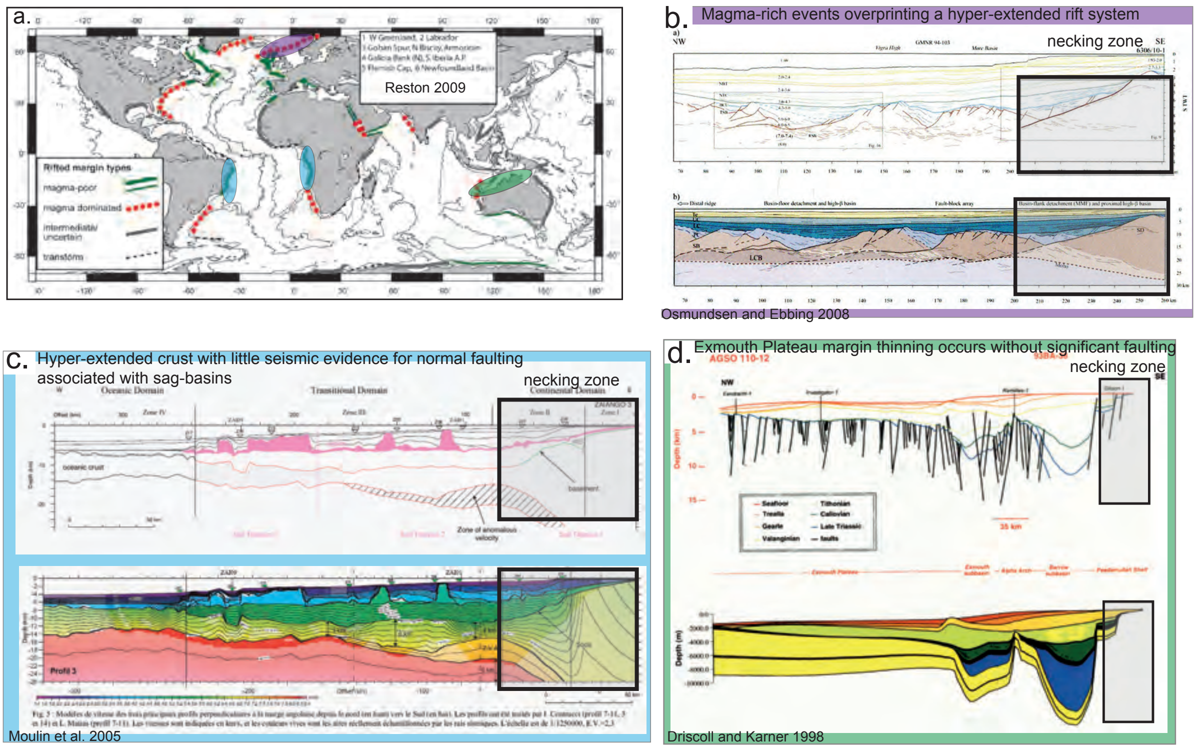


Fig. I. 4: (a) Carte géographique représentant les différents types de marge : peu magmatique, magmatique et oblique (Reston 2009). (b) La marge passive Norvégienne, interprétation de la ligne sismique GMNR-94-104 (Osmundsen et Ebbing 2008). Il est important de noter que la présence de croûte continentale très amincie se trouve également dans les marges magmatiques. (c) La marge passive Angolaise, en haut : Profil 3 de ZaïAngo, (échelle verticale en profondeur). En bas, modèles de vitesse de la marge Angolaise. Cette marge montre une croûte continentale dans les parties distales de la marge très amincie ne montrant pas de structures extensives capable d'accommoder cette extension (Moulin et al. 2005). (d) La marge nord-ouest Australienne, section sismique (en profondeur) montrant l'architecture du bassin de Exmouth et de la partie Sud du bassin de Barrow. Il est important de noter que l'importante subsidence observée au Tithonien-Valenginien n'est pas accompagnée par une déformation cassante visible dans la croûte supérieure (Driscoll et Karner 2008).

proximales (Fig. 5) (Lau et al 2006). La géométrie des zones proximales contraste avec celle observée dans les parties distales et la TOC comme cela est explicité dans le paragraphe précédent. En effet, dans la marge distale, la croûte se trouve amincie de 30 à 10 km d'épaisseur, puis grâce au jeu de faille de détachement elle est amincie de 10 à 0 km conjointement à l'exhumation de manteau subcontinental (Figs. 3, 5). Dès lors, l'objectif majeur de mon étude est de comprendre comment et quand une croûte continentale équilibrée de marge proximale est relayée par une croûte extrêmement amincie dans les marges distales (≤ 10 km). Ces domaines distincts de marge sont caractérisés par l'absence de structure visible sur coupe sismique pouvant accommoder cet amincissement extrême. La réponse à cette question ne se trouve pas dans la marge distale sensu-stricto qui a déjà subi cet amincissement extrême. En revanche, la transition entre la marge proximale et distale est caractérisée en coupe sismique par une remontée rapide du Moho avec un angle pouvant atteindre 35° (Fig. 5) (Lau et al. 2006). Cette zone appelée zone d'étranglement (ou «necking zone») enregistre le passage d'une croûte de 30 km à 10 km (Fig. 5). Cette «necking zone» est identifiable dans de nombreuses marges passives peu magmatiques actuelles (marge Ibérique, Atlantique Sud) (Figs. 4, 5) (e.g. Péron-Pinvidic et Manatschal 2009 ; Contrucci et al. 2004 ; Moulin et al. 2005). Elle est également observable dans les marges passives magmatiques lorsque l'extension lithosphérique qui amincit la croûte jusqu'à une épaisseur de 10 km est antérieure au principal événement magmatique (Fig. 4) (Osmundsen et Ebbing 2008).

Ce travail est articulé autour de plusieurs questions et thématiques majeures :

a. Quels processus pendant le rifting contrôlent l'extrême amincissement crustal entre la marge proximale montrant une épaisseur crustale quasi normale et la marge distale avec une épaisseur crustale voisine de 10 km ? (Publication 2-3)

b. Quelle est la répartition spatiale des différents niveaux de croûte dans la zone d'étranglement ainsi que dans la marge distale ? (Publication 2)

c. Quelle est l'évolution tectonique, spatiale et temporelle du rifting à l'échelle des marges passives peu magmatiques ? (Publication 3)

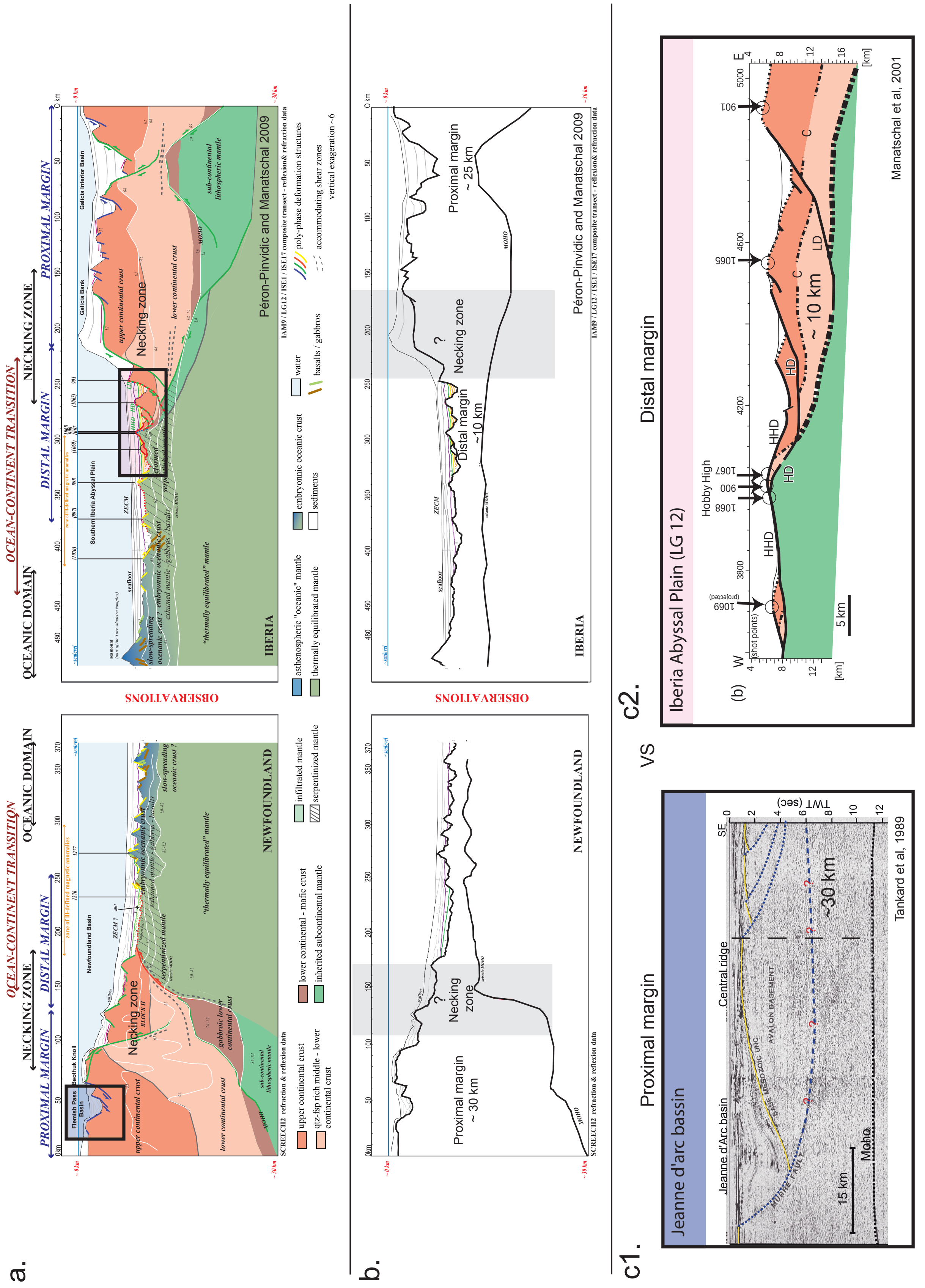
3. Où et Comment répondre à ces questions ?

Les donnée géologiques et géophysiques sur les marges passives actuelles ne permettent pas de répondre à aux questions posées précédement. En effet, les données de sismiques réfractions ne permettent d'avoir qu'un contrôle très partiel de l'architecture crustale. L'identification des différents niveaux crustaux sont soumis au modèle de vitesse sismique. Or, les roches peuvent changer de vitesses sismiques sous l'effet de changements des conditions P-T, d'altération, de la fracturation. Par exemple une roche de croûte inférieure avec une relative « haute » vitesse sismique pourra, sous l'effet de l'exhumation et de processus de surface, avoir une vitesse comparable à la croûte supérieure. Dès lors, l'identification précise des roches de croûte supérieure, moyenne, et inférieure reste une question très délicate.

Ces incertitudes sont augmentées par le nombre relativement faible de forages dans les parties distales des marges (12 forages). De plus aucun forage ODP n'a à l'heure actuelle été fait dans la «necking zone» (Fig. 5). Tous ces éléments rendent les sites des marges passives actuelles peu favorables pour fournir les arguments de réponses aux questions concernant l'amincissement crustal.

Afin de tenter de répondre à ces questions, ma thèse est donc logiquement focalisée sur le domaine Téthysien Alpin qui possède les reliques des marges passives Européenne et Adriatique (Fig. 6). Ce domaine a fait l'objet d'études depuis plus d'un siècle et bénéficie ainsi d'une base de données inégalée par rapport à d'autres systèmes fossiles de marge peu magmatique. Grâce aux nombreux travaux antérieurs (e.g. Argand 1916 ; Trümpy 1980 ; Schmid et al. 1996, 2004 et références citées), ce domaine possède une cartographie détaillée ainsi qu'une stratigraphie très bien contrainte (e.g. Lemoine et al. 1986 ; Bertotti et al. 1993 et références citées). L'histoire Phanérozoïque des Alpes a fait l'objet de nombreuses études qui permettent un contrôle au premier ordre de l'évolution tectonique Alpine. Tous ces paramètres font du système Téthysien Alpin un laboratoire naturel de premier plan pour la compréhension des processus d'extension lithosphérique. Mon étude est plus précisément centrée sur le domaine Adriatique car celui-ci n'a été que peu affecté par la réactivation Alpine en compression (Figs. 6, 7). En effet ce domaine a été localisé au toit de la subduction du domaine Téthysien et a échappé à une superposition métamorphique très intense dans des faciès de haute pression préservant ainsi les structures pré-Alpines.

L'étude a pour base un travail de terrain et de géologie structurale visant à carter et comprendre les différentes géométries issues des diverses phases de déformation visibles sur le terrain. Ces analyses sont complétées par l'établissement de nouvelles coupes Alpines ainsi que des coupes rétro déformées de la zone d'étude. Avec l'apport des précédents travaux une


a.
PROXIMAL MARGIN
NECKING ZONE
DISTAL MARGIN
OCEAN-CONTINENT TRANSITION
OCEANIC DOMAIN
OBSERVATIONS
NEWFOUNDLAND
IBERIA
Necking zone
Péron-Pinvidic and Manatschal 2009
b.
Proximal margin ~ 30 km
Necking zone
Proximal margin ~ 25 km
Distal margin ~10 km
c1.
Jeanne d'arc bassin
Proximal margin
~30 km
15 km
Moho
Tankard et al, 1989
vs
c2.
Iberia Abyssal Plain (LG 12)
Distal margin
~ 10 km
5 km
Manatschal et al, 2001

carte géologique générale de la zone est établie. L'analyse de terrain est couplée aux données pétrologiques et thermochronologiques par la méthode de datation Ar/Ar obtenues sur biotite, muscovite et amphibole.

4. Le choix de la zone d'étude

Le choix de la zone d'étude est un point important. Aucune «necking zone» n'a été identifiée dans les marges passives fossiles. Cependant, dans la marge Adriatique des reliques de marge proximale (Froitzheim 1988 ; Conti et al. 1994), distale (Froitzhein et Eberli 1990, Manatschal et Nievergelt 1997) et de TOC (Desmurs et al. 2001) ont été identifiées récemment (Fig. 7, 8). Associée à une faible déformation Alpine, la paléo-marge Adriatique représente une opportunité unique pour étudier l'architecture d'une marge passive très peu réactivée dans une chaîne de collision comme les Alpes. Les reliques de cette marge affleurent principalement dans les Alpes du Sud près de la région de Turin, ainsi qu'au Sud –Est de la Suisse et au Nord de l'Italie (Figs. 6, 7). La marge Adriatique dans les Alpes du Sud a fait l'objet de nombreuses études essentiellement sur la zone d'Ivrée qui montre des reliques de croûte inférieure pré-rift qui ont été exhumées (revue par Handy et al. 1999 et références citées). Cependant, les conditions très défavorables d'affleurement et l'impossibilité de disposer d'une continuité structurale entre les différentes parties de la marge en font une zone peu appropriée pour mon étude. J'ai donc préféré retenir les sites des reliques de la marge Adriatique exposées au Sud-est de la Suisse (canton des Grisons) et au Nord de l'Italie (en région Lombarde dans la province de Sondrio) dans les nappes Austroalpines et Sud Pénnique. La particularité de cette zone découle de sa position dans la chaîne Alpine. Cette zone affectée par deux systèmes orogéniques distincts, au Crétacé et Tertiaire, a cependant remarquablement échappé à une déformation trop importante.

Les nappes Austroalpines et Sud-Pénnique se décomposent en plusieurs ensembles d'unités qui se sont retrouvées superposées par la première phase de compression. Celle-ci est caractérisée par une succession de chevauchements. Ceci a eu pour incidence l'empilement des différentes unités de la marge les unes sur les autres. Ainsi, de l'Est vers l'Ouest les unités d'affinité continentale précédent des unités « océaniques ». Le sens de transport des chevau-

Fig. I. 5: (a) Reconstruction des marges conjuguées de Terre Neuve et Ibérie (Péron-Pinvidic et Manatschal 2009). (b) croquis de la reconstruction précédente montrant les variations de l'architecture crustale à travers les parties proximales vers les parties distales de la marge. Il est important de noter que les changements d'épaisseurs les plus importants se situent dans la necking zone. (c1) Coupe sismique réflexion à travers le bassin de Jeanne d'Arc (Tankard et al. 1989) montrant que la déformation le long de faille normale à fort pendage n'affecte que la partie supérieure de la croûte sans amincissement visible. (c2) Profil sismique de la coupe Lusigal 12 (Manatschal et al. 2001) montrant que la déformation dans les parties les plus distales des marges est contrôlée par des failles de détachement à faible pendage impliquant l'exhumation de croûte supérieure et de manteau subcontinental.

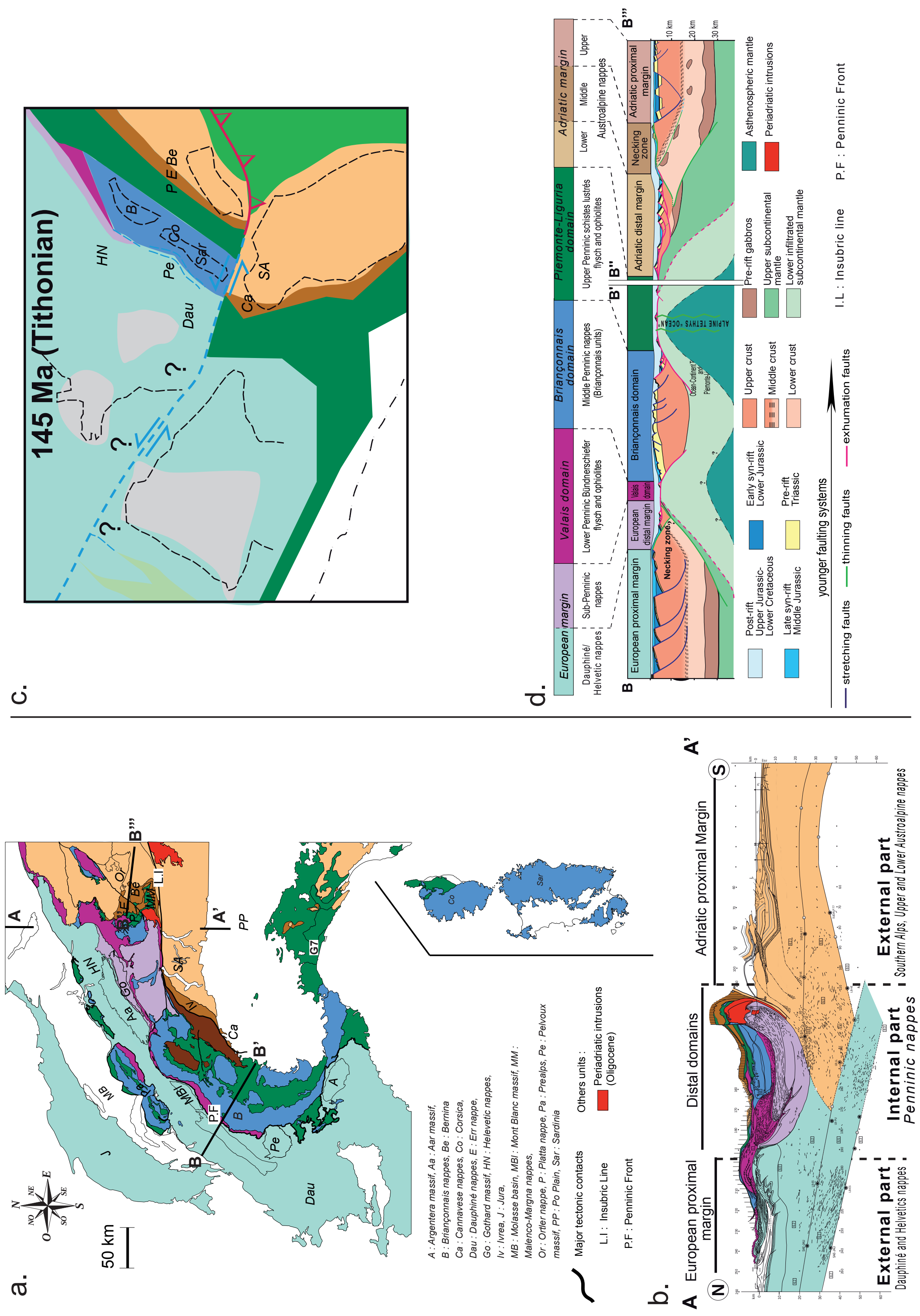


Fig. I. 6: (a) carte géologique des Alpes montrant la position paléogéographique des principales unités tectoniques représentant les reliques des marges passives de la Téthys Alpine (modifiée d'après Schmid et al. 2004). (b) Coupe sismique NFP 20 montrant la répartition des principales unités Tectoniques Alpines représentées dans la figure (a) (modifié d'après Schmid et al. 1996). (c) Reconstruction paléogéographique des différentes unités des marges de la Téthys Alpine au Tithonien. (d) Coupe reconstruit montrant l'architecture des marges de la Téthys Alpine à la fin du Jurassique (Tithonien).

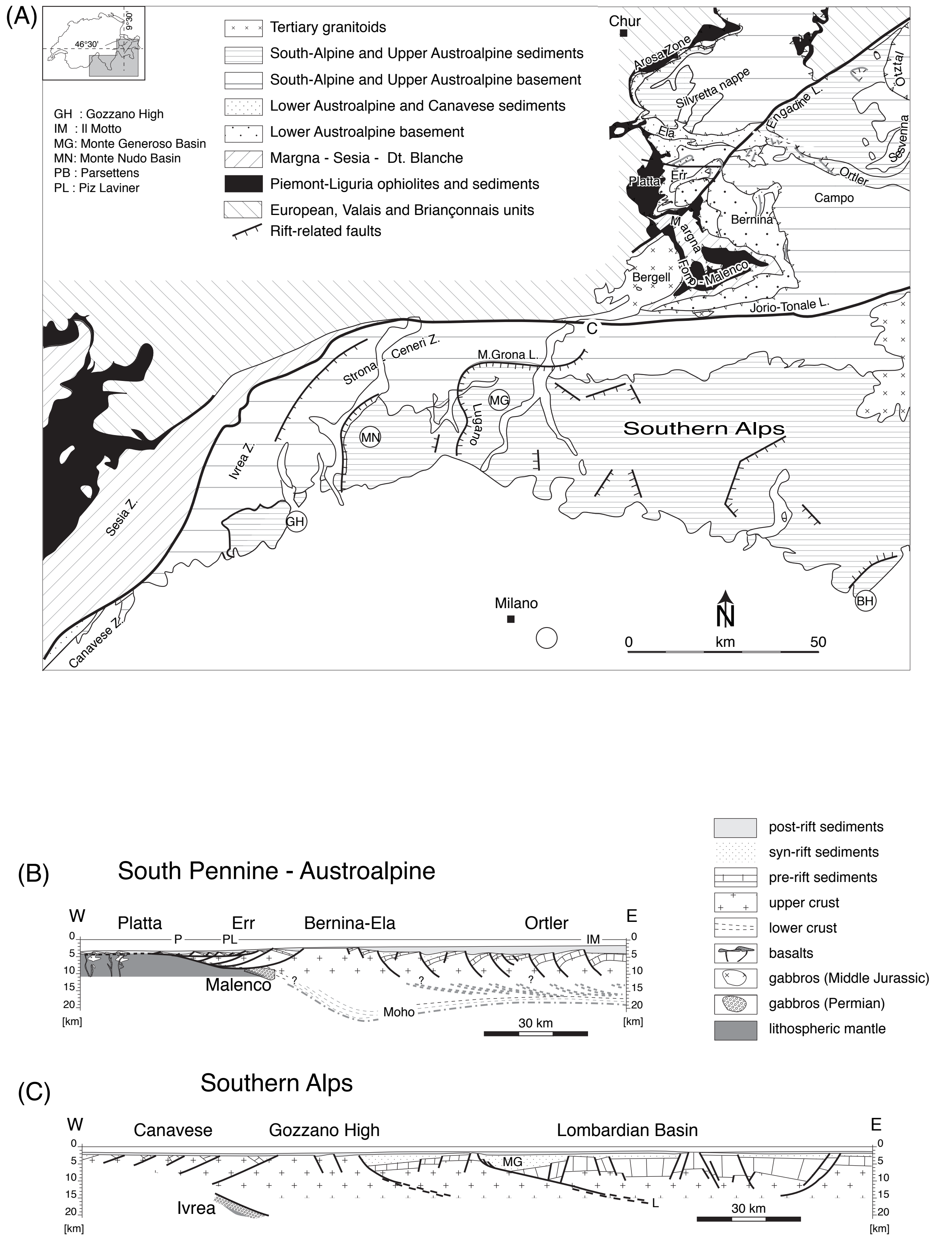


Fig. I. 7: (a) Carte géologique des Grisons et Sud des Alpes, d'après Bernoulli et al. (1990), Müntener et Hermann (2001). (b) Reconstruction de la géométrie de la marge Adriatique préservée dans les Grisons au Jurassique (Manatschal et Bernoulli 1999). (c) Reconstruction de la marge Adriatique préservée dans les Alpes du Sud au Jurassique (Bertotti et al. 1993) chements est parallèle au profil de la marge. L'empilement s'effectue dans l'ordre suivant : de l'unité continentale ainsi que de la plus haute structuralement parlant, à l'unité la plus océanique et de niveau structural le plus bas :

1. *L'unité d'Ortler* (Fig. 8). Celle-ci se compose d'une épaisse couverture mésozoïque principalement composée de dépôts Triasiques et Jurassiques. La présence de failles normales à fort pendage qui se sont initiées au début du rifting produisant une géomètrie de bloc basculé a été établie, et cette zone a été interprétée par de nombreux auteurs (i.e. Froitzheim 1988) comme une relique d'une ancienne marge proximale formée pendant le rifting Jurassique.
2. *Les unités de Bernina-Campo-Grosina* (Fig. 8). Ces unités n'ont été que peu étudiées et restent à l'heure actuelle mal décrites et mal comprises. En conséquence, aucune position paléogéographique précise ne leur a été assignée.
3. *L'unité d'Err* (Fig. 8). Froitzheim et Eberli (1990) ont identifé des failles de détachement à faible pendage dans la croûte continentale. Le dépôt de sediments syn-rift avec un onlap de 10 à 20° ainsi que la présence de clastes de détachement dans les sédiments syn-rift permettent de confirmer que cette faille a été active pendant le rifting. Ces observations ont été confirmées ultérieurement par Manatschal et Nievergelt (1997) qui ont identifié cette zone comme étant une marge distale.
4. *L'unité de Platta* (Fig. 8). Elle est composée essentiellement de manteau sub-continental serpentinisé. Le toit du manteau est coiffé par une faille de détachement qui est scellé par des sédiments post-rift. Manatschal et Nievergelt (1997), ainsi que Desmurs et al. (2001) ont interprété cette unité comme étant une ancienne TOC.

La description plus précise de diverses unités ainsi que leurs relations tectoniques et évolutives sont données en détail dans la publication 1.

L'étude est focalisée sur les unités de Bernina-Campo-Grosina. En effet, ces unités sont situées à l'interface entre les domaines paléogéographiques de la marge proximale et de la marge distale. Ainsi, la position des unités de Bernina-Campo-Grosina est dans une position similaire à la «necking zone». Ma thèse se propose d'analyser ces unités afin de décrire et comprendre la tectonique Alpine, puis dans un deuxième temps de reconnaître les structures liées au rifting.

La description ainsi que la compréhension des structures compressives Alpines sont un pré-requis essentiel pour caractériser les structures pré-Alpines et plus particulièrement celles dues au rifting. Ceci est développé dans la publication 1, ayant pour but de caractériser la géométrie pré-Alpine de la zone d'étude :

e. Comment les structures de rifting sont elles réactivées pendant la phase compressive Alpine ? Quel impact ont ces structures issues du rifting sur la géométrie des structures Alpines ? (Publication 1)

f. Quelle est le rôle de l'héritage structural aussi bien sur le contrôle du rift ainsi que sur l'orogenèse Alpine ? (Publication 1-3)

g. Quelles sont les implications d'une meilleure connaissance des processus d'extension lithosphérique sur les mécanismes compressifs de l'orogène ? (Publiction 1-3)

5. La démarche et le plan de la thèse

Le mémoire de thèse comporte deux parties distinctes concernant deux différentes échelles. La première partie est dédiée à la description de la tectonique Alpine et à la rétrodéformation des unités de Bernina-Campo-Grosina appartenant aux nappes Austroalpines dans le Sud-est de la Suisse et le Nord de l'Italie.

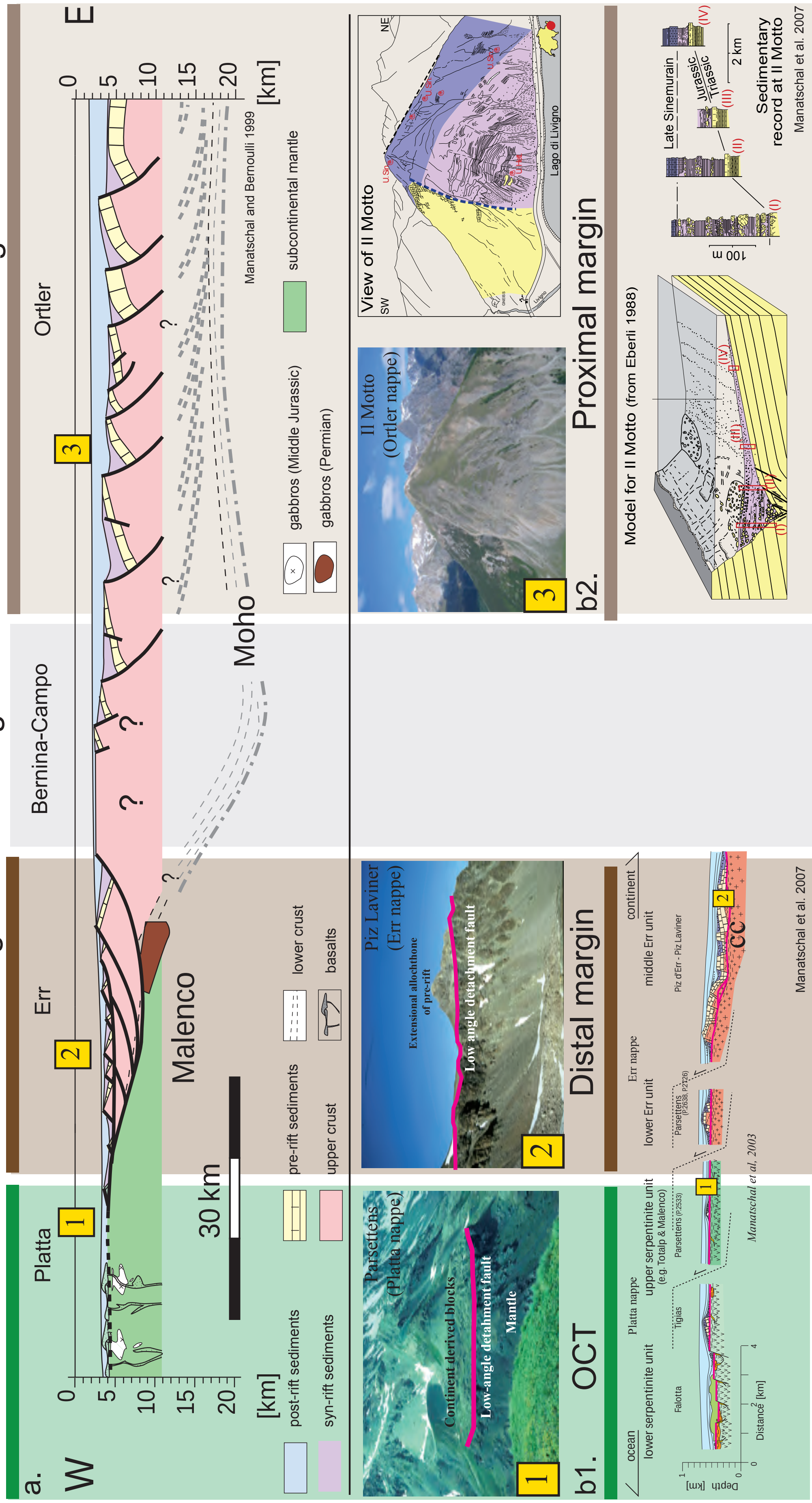
Cette réflexion est développée dans les publications 1 et 2.

La publication 1 comporte de nouveaux éléments pour la compréhension de l'histoire de la tectonique Alpine des nappes Austroalpines et Sud Pénnique. La deuxième partie, expose l'identification et la reconstruction Alpine des unités de Bernina-Campo-Grosina. Cette réflexion permet de faire le lien entre les structures pré-Alpines et plus particulièrement celles issues du rifting, avec la réactivation Alpine. L'importance de la réactivation lors des processus de collision est discutée.

Publication 1: "Revisiting the Austroalpine nappes in SE-Switzerland and N-Italy: new constrains on the Alpine evolution and implication for strong pre-Alpine inheritance" International Journal of Earth Science, volume spécial en l'honneur de Rudolf Trümpy.

La publication 2 est dédiée à l'analyse de la «necking zone». Cet article présente une description des structures extensives issues du rifting pouvant accommoder l'étirement lithosphérique. L'objectif est d'apporter des contraintes géologiques fortes permettant de tester les modèles actuels.

Publication 2: "How does the continental crust thin during rifting in magma-poor rifted margins: evidence from the Bernina/Campo/Grosina units in the Central Alps (SE-Switzerland and N-Italy) and implications for present-day rifted margins" Tectonics.



a.
W
E
OCT
Distal margin
Necking zone
Proximal margin
Platta
Err
Bernina-Campo
Ortler
Malenco
Moho
30 km
[km]
0
5
10
15
20
post-rift sediments
syn-rift sediments
pre-rift sediments
upper crust
lower crust
basalts
gabbros (Middle Jurassic)
gabbros (Permian)
subcontinental mantle
Manatschal and Bernoulli 1999
Parsettens (Platta nappe)
Continent derived blocks
Low-angle detachment fault
Mantle
Piz Laviner (Err nappe)
Extensional allochthone of pre-rift
Low angle detachment fault
Il Motto (Ortler nappe)
View of Il Motto
SW
NE
Lago di Livigno
b1.
OCT
Distal margin
b2.
Proximal margin
ocean
lower serpentinite unit
Platta nappe
upper serpentinite unit (e.g. Totalp & Malenco)
Err nappe
lower Err unit
middle Err unit
continent
Falotta
Tigias
Parsettens
Piz d'Err - Piz Laviner
CC
Depth [km]
Distance [km]
Manatschal et al. 2003
Manatschal et al. 2007
Model for Il Motto (from Eberli 1988)
Late Sinemurain
Jurassic
Triassic
2 km
100 m
Sedimentary record at Il Motto
Manatschal et al. 2007

La deuxième partie du mémoire concerne l'ensemble de la chaîne Alpine (essentiellement dans les Alpes de l'Ouest et Centrale). L'objectif a été de reconsidérer l'ensemble du système de la Téthys Alpine et ainsi de proposer un modèle global pour l'évolution des marges passives peu magmatiques. Cette partie a une approche plus globale, portant à la fois sur les mécanismes de l'extension à l'échelle de la lithosphère et sur pour la compréhension de la formation de la chaîne Alpine. Les résultats sont discutés dans la publication 3.

La publication 3 concerne une étude menée sur une échelle plus large, celle de la chaîne Alpine dans son ensemble. En effet, les nombreux travaux accomplis sur cette chaîne donnent l'opportunité de synthétiser les données, pour en retracer l'évolution durant le rifting. Cette synthèse tient compte des conditions présentes pendant la période de pré-rift, puis pendant le syn- et post- rift. L'implication de la structuration de la lithosphère pendant le rifting pour la géométrie de l'orogène Alpine est évaluée.

Publication 3: "Unravelling the interaction between tectonic and sedimentary processes during lithospheric thinning in the alpine tethys margins" International Journal of Earth Science, volume spécial pour les 100 ans du journal.

Fig. I. 8: (a) Reconstruction de la marge Adriatique préservée dans les Grisons (Manatschal et Bernoulli 1999) (b1) Reconstruction palinspastique de la marge distale Adriatique à travers la section d'Err-Platta dans les Grisons. La déformation lors du rifting Jurassique dans ces unités est contrôlée par des failles de détachement à faible pendage au toit de la croûte contientale (photo 2) ainsi que toit du manteau subcontinental (photo 1) (Manatschal 2007 et références cités). (b2) (Photo 3) Vue d'Il Motto représentant un bloc basculé issu du rifting Jurassique dans les parties proximales de la marge Adriatique. La faille normale à fort pendage est scellée par les dépôts du Sinémurien supérieur (Manatschal 2007 et références cités). Il est important de noter que le style de déformation extensive entre les parties proximales et distales des marges passives est très différent. Les unités de Campo-Grosina-Bernina se situent à l'interface entre ces deux domaines.

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DEUXIÈME PARTIE

Le chapitre premier présente la géologie des nappes Austroalpines qui préservent les reliques de la marge fossile Adriatique dans les Sud-Est de la Suisse et le Nord de l'Italie. La compréhension de la géologie Alpine de la zone d'étude permet de proposer une restauration de la géométrie de la marge passive Adriatique à la fin du rifting. Ce chapitre résulte essentiellement d'un travail de terrain effectué pendant les étés 2007, 2008, 2009. Il a consisté en des travaux de cartographie, de géologie structurale et en l'établissement de coupes géologiques à travers la zone étudiée.

Ce chapitre est composé de trois parties :

La première partie a pour but de présenter : 1) une nouvelle classification des nappes décrites, 2) la stratigraphie, 3) l'histoire paléozoïque du socle et 4) la synthèse des différents événements de déformations lors de la réactivation de la marge Adriatique créant les nappes Austroalpines. Dans cette première partie qui utilise avant tout les données des travaux antérieurs, de nouvelles observations complémentaires et des interprétations sont proposées pour caractériser et définir le contexte tectonique lors du dépôt de certaines formations stratigraphiques ainsi que la succession de différentes phases de déformations.

La deuxième partie est focalisée sur les nappes de Campo, Grosina, Bernina. Cette partie expose les nouvelles observations effectuées sur la zone. De nouvelles cartes et coupes Alpines sont proposées, permettant ainsi la distinction entre des structures reliées à la réactivation de la marge Adriatique et celles issues du rifting.

La troisième partie est une discussion des résultats. Il découle de cette étude que la nappe de Bernina préserve les reliques d'une ancienne marge distale tandis que les nappes de Campo/Grosina montrent les reliques d'une « necking zone » ou zone d'étranglement. Un accent tout particulier est mis sur le contrôle des structures héritées du rifting pour la réactivation de la marge Adriatique de la fin du Crétacé jusqu'au Tertiaire. Une coupe restaurée de la zone d'étude est présentée montrant l'importance de l'héritage structural ainsi que la complexité de la géométrie de la marge Adriatique à la fin du rifting.

Cette partie du travail de thèse fait l'objet d'une publication dans un volume spécial du journal « International Journal of Earth Sciences» (IJES) en l'honneur de Rudolf Trümpy. Le chapitre présenté dans le mémoire est une version longue de la publication, complétée par des données bibliographiques étendues et la présentation détaillée des nouvelles observations acquises.

Publication 1: submitted to IJES for the Rudolf Trümpy volume

REVISITING THE AUSTROALPINE NAPPES IN SE-SWITZERLAND AND N-ITALY: NEW CONSTRAINS ON THE ALPINE EVOLUTION AND IMPLICATION FOR STRONG PRE-ALPINE INHERITANCE

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ABSTRACT

The Austroalpine nappe systems in SE-Switzerland and N-Italy preserve remnants of the Adriatic rifted margin. Despite numerous previous studies, the structure and architecture of the Campo, Grosina/Languard and Bernina nappes remain unclear and controversial. Based on new maps and cross sections, we propose that this complexity is largely inherited from Jurassic rifting. Considering the structure and stratigraphic evolution of the Adriatic rifted margin we propose a new classification of the Austroalpine domain, which includes Upper, Middle and Lower Austroalpine nappes. In this paper we describe the Alpine structures as well as the pre-Alpine rift-related geometry of the Lower (Bernina) and Middle (Campo, Grosina/Languard) Austroalpine nappes in order to propose a restoration of the position of these nappes along the former margin, as well as to understand the formation and emplacement of the nappes during initial reactivation of the Alpine Tethyan margin. We show that the Campo and Grosina/Languard nappes can be interpreted as remnants of a former necking zone that comprised pre-rift upper and middle crust. These nappes were juxtaposed against the Mesozoic cover of the Bernina nappe during Jurassic rifting. We find evidence for low-angle detachment faults and extensional allochthons in the Bernina nappe, similar to those previously described in the Err nappe and explain their role during subsequent reactivation. Our observations from the Bernina-Campo-Grosina/Languard nappes reveal a strong control of rift-related structures during the subsequent Alpine reactivation on all scales of the former distal margin.. Two zones of intense deformation can be mapped, referred to as the Albula-Zebru and Lunghin-Mortirolo movement zones. These zones have been reactivated during Alpine deformation and can not be described as simple monophase faults or shear zones. We propose a tectonic model for the Austroalpine nappe systems that links inherited rift-related structures with the present-day Alpine structures. In conclusion, we believe that apart from the direct regional implications, the results of this paper are of general interest in understanding the control of rift structures during reactivation of distal rifted margins.

Keywords: Austroalpine, Bernina/Campo, Alpine tectonics, rifting, tectonic inversion

1. Introduction

The Austroalpine nappe systems in SE Switzerland and N-Italy have been intensely studied for more than a century (e.g. Steinman 1925, 1927; Staub 1946, Trümpy 1975). They preserve one of the best-documented ancient rifted margins (e.g. Froitzheim and Eberli 1990; Bernoulli et al, 1990; Manatschal and Nievergelt 1997) and represent therefore a key area for the investigation of the reactivation of rift-related structures within a collisional orogen. Numerous stratigraphic (e.g. Eberli 1988; Furrer et al. 1985 and references therein), petrological (e.g. Trommsdorff et al. 1993; Müntener et al. 2000; Hermann et al. 2001) and structural (e.g. Froitzheim et al. 1994 and references therein) studies unraveled the complex Alpine and pre-Alpine deformation history recorded in the Austroalpine nappe stack. The aim of this paper is to revisit the Austroalpine nappe systems in order to investigate the relation between pre-Alpine and Alpine structures. Key questions addressed in this paper are: How is the structural architecture of these nappes controlled by inherited structures? What is the relation between upper, middle and lower crustal and mantle rocks found in the Austroalpine and Upper Penninic nappe systems, and when and how were these different levels juxtaposed against each other? We will show that in order to resolve and understand the Alpine structures and deformation history in the Austroalpine nappe systems it is vital to understand the pre-Alpine history.

In this paper we present a new geological map of the Austroalpine nappe systems in Grisons and N-Italy, propose and discuss new cross sections to redefine the paleogeographic domains from which these nappes derived. These reconstructions will be used to discuss the role of inheritance during the reactivation of the distal parts of the Adriatic rifted margin and to propose a restoration of the Bernina, Campo and Grosina/Languard nappes, showing that these nappes record the missing link between the more proximal and distal parts of the reconstructed former Adriatic rifted margin.

2. Regional geological setting

The study area discussed in this paper is located in SE-Switzerland and N-Italy and comprises the Upper Penninic and Austroalpine nappe systems (Figs. 1, 2). These nappes sampled parts of the northwestern Adriatic margin and were located at the interface between two discrete orogenic events related respectively to the closure of the Neotethys (northern part of the Meliata-Vardar oceanic domain; e.g Channel and Kozur 1997) and Alpine Tethys (e.g Froitzheim et al. 1996) (Fig. 1). The nappe stack was emplaced in an external part of the first orogen resulting from the closure of the northern Meliata-Vardar domain in Late Cretaceous time and was overprinted during the closure of the Alpine Tethys domain during Latest Cretaceous to Late Eocene (Figs. 1 and 2). During the subsequent continental collision between Europe and Adria (latest Eocene to Miocene), the previously formed nappe stack was affected by N-S shortening, resulting in a N-vergent fold and thrust belt in the north and a S-vergent back-folding to the south (Fig. 2). The nappes studied in this paper were situated in a neutral zone within the Alpine orogen (Fig. 2b). This position explains the weak Alpine tectonic overprint during the Cenozoic Alpine collision and the excellent preservation of rift-related structures in SE-Switzerland (Fig. 2).

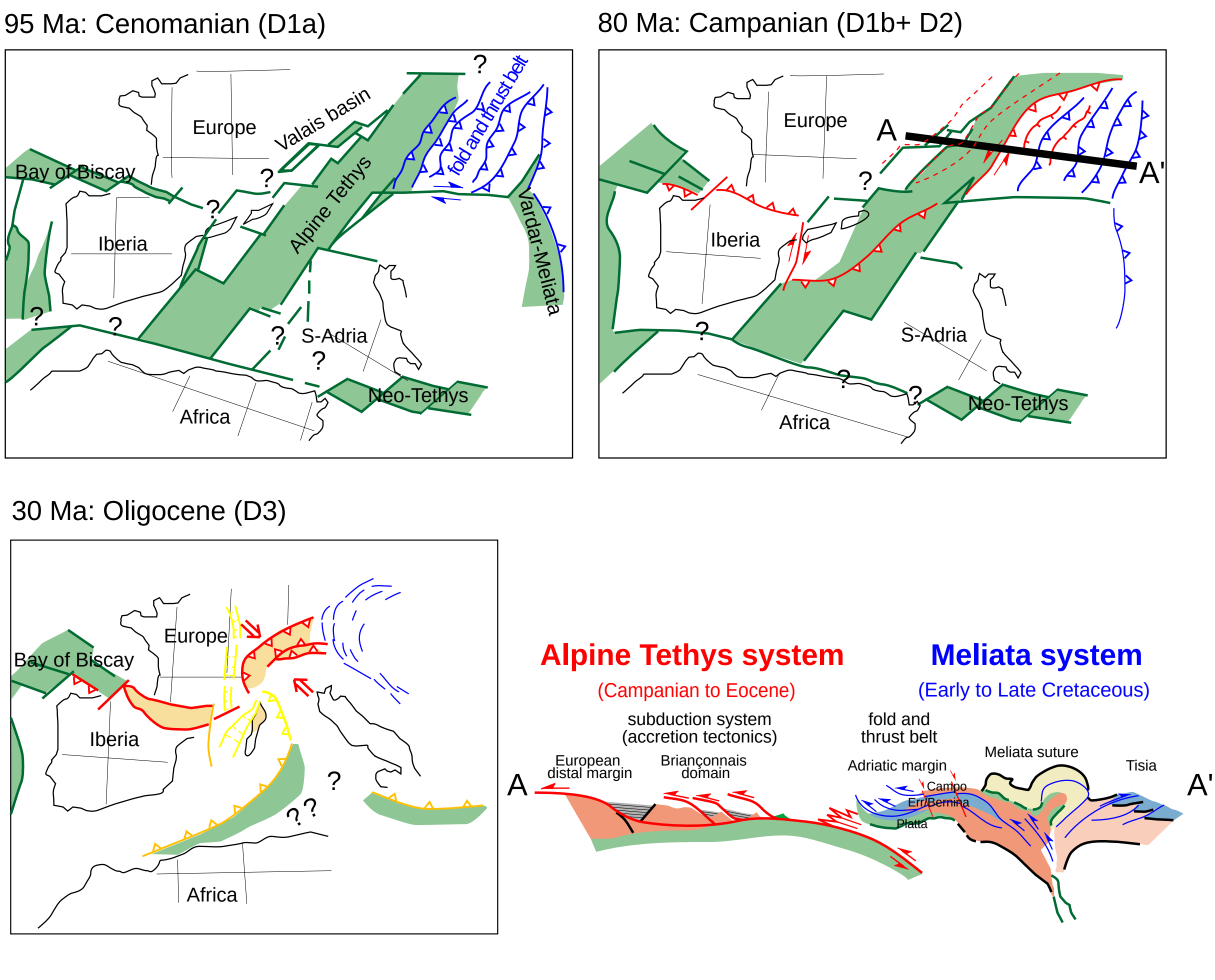


Fig. II. 1: Paleogeographic maps of the Alpine domain for the Cenomanian (95Ma), the Late Cretaceous (Campanian 80Ma) and the Oligocene (30Ma). The section in the lower right corner shows the Adriatic-European margins during the Campanian corresponding to the transition from the final collision of the northern Vardar-Meliata system to the onset of subduction in the Alpine Tethys domain (modified from Manatschal and Müntener 2009).

The reconstruction presented in this paper will focus mainly on the restoration of a section across a central part of the study area (Fig. 3). Deformation to the north, along the base of the Upper Austroalpine Ortler-Ela nappes is complex and includes late, out-of-sequence deformation structures. To the south, the Sella-Margna and Malenco nappes are strongly overprinted by Alpine tectonic of the previously formed nappe stack. Further to the south, the nappes are affected by back thrusting and movements along the peri-Adriatic fault system (e.g. Insubric line; Schmid et al. 1989). To the west, the Upper Penninic nappe system is truncated by the Turba shear zone, which is an Eocene normal fault (Nievergelt et al. 1996). The footwall of this shear zone is made of Middle and Upper Penninic nappe systems, which have been highly deformed and metamorphosed by the Cretaceous and subsequent Tertiary collision.

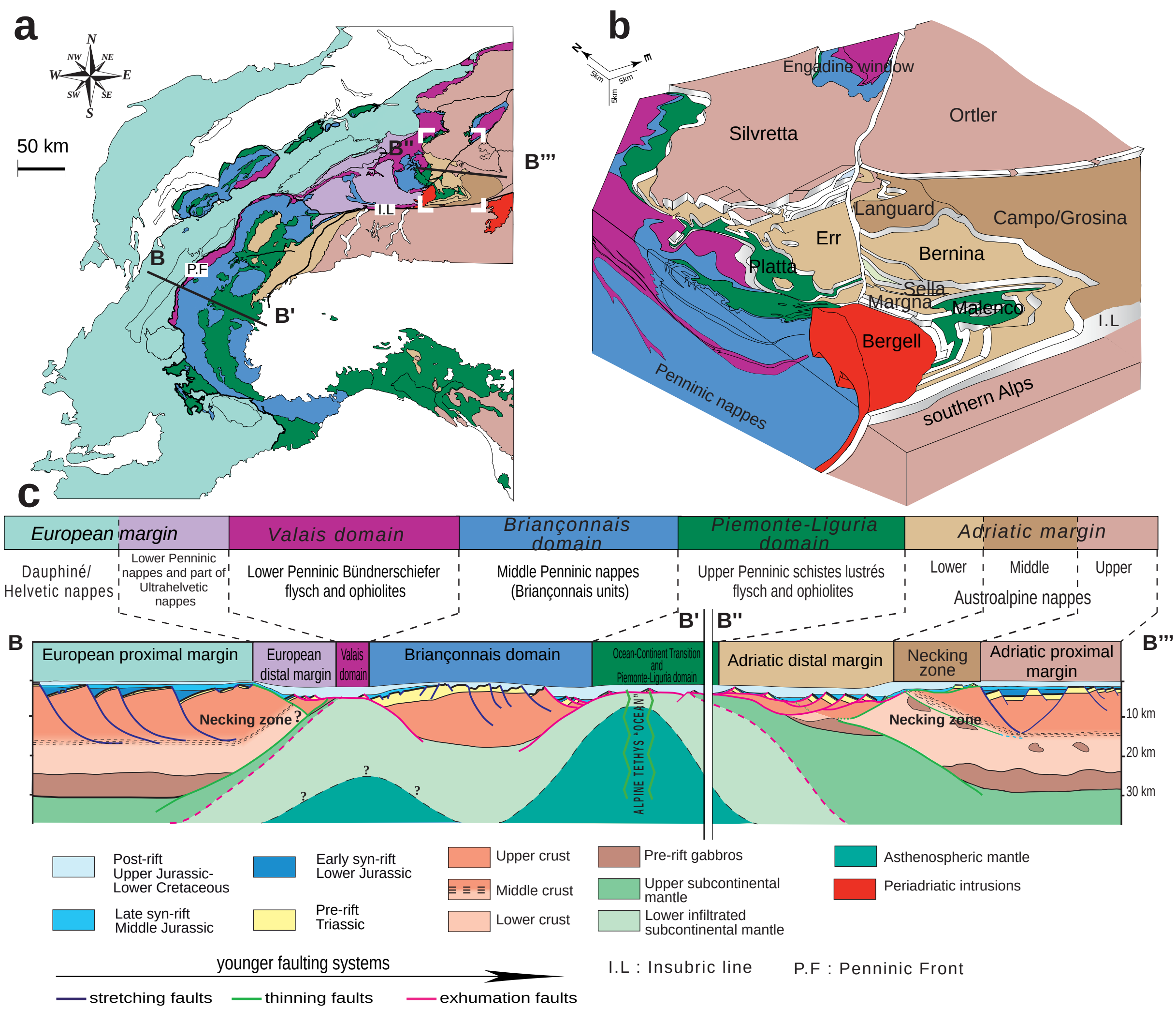


Fig. II. 2: (a) Tectonic overview map of the Alps showing the distribution of the main paleogeographic domains (modified after Schmid et al. 2004). (b) Schematic block diagram of SE-Switzerland and N-Italy showing the position of the Austroalpine and Penninic nappes and associated units (modified after Froitzheim et al. 1994). (c) Cross section across the Alpine Tethys margins at the end of rifting and before onset of Alpine convergence in Late Cretaceous time (modified from Mohn et al. (2010)). For the location of the cross section see tectonic map above (a).

The Austroalpine Bernina, Campo, and Grosina/Languard nappes have been studied and mapped over more than a century by Austrian, Swiss and Italian geologists (Cornelius (1932, 1935, 1950); Staub (1946, 1964); Spillmann and Trommsdorff (2005); Trümpy (1980); Froitzheim et al. (1994), Peters (2005, 2007). Despite the careful mapping of these previous workers, no detailed and coherent map of the entire area located between the Valtellina, Albula and Julier valleys exists. In this paper, we present a new geological map (Fig. 3) and constructed sections across this zone; we propose a new subdivision of the Austroalpine nappe systems into Upper, Middle and Lower Austroalpine nappes based on paleogeographic considerations of the Adriatic rifted margin that was situated next to the Alpine Tethyan ocean in Jurassic time (Mohn et al. 2010). The term "Middle Austroalpine" has been the subject of many debates. It was first defined by Staub (1920) and subsequently established by Tollmann (1959, 1977) based on the subdivision of the Austroalpine basement and its Early Mesozoic sedimentary cover according both to their present position in the nappe pile in the Eastern Alps (Upper, Middle, Lower) as well as to their presumed location within the Triassic continental margin of Adria. This older margin developed prior to opening of the Neotethyan Ocean to the southeast and well before the Jurassic opening of Alpine Tethys to the northwest of Adria.

However, recent studies of Schuster and Frank (1999) and Schuster et al. (2001) on the polymetamorphic evolution of the Austroalpine basement have challenged Tollmann's classification and concluded that the term "Middle" Austroalpine should be avoided on Early Mesozoic paleogeographic grounds. They propose that Upper and Lower Austroalpine units formed during Triassic rifting, then were juxtaposed during Jurassic strike-slip faulting. In this paper, however, we reintroduce the term "Middle Austroalpine" in a different paleogeographic context that is based on the architecture of the Jurassic rifted margin of Adria that developed just before spreading of the Piemont branch of the Alpine Tethyan Ocean. Indeed our distinction of Upper, Middle and Lower Austroalpine nappes correlates directly with a proximal margin, a necking zone and a distal margin in Jurassic time, similar to the architecture found on modern rifted margins (Mohn et al. 2010). The necking zone represents the transition between the proximal and distal parts of rifted margin where the continental crust thins from 30 km to less than 10 km (for a modern analogue, see Fig. 2 in Péron-Pinvidic and Manatschal 2009; Osmundsen and Ebbing 2008). We will show that the Middle Austroalpine nappe systems defined in this work (Campo, Grosina/Languard) are distinct from the Upper and Lower Austroalpine nappe systems and preserve characteristics of a former necking zone. Consequently, in this work, we distinguish between the following systems (Fig. 3):

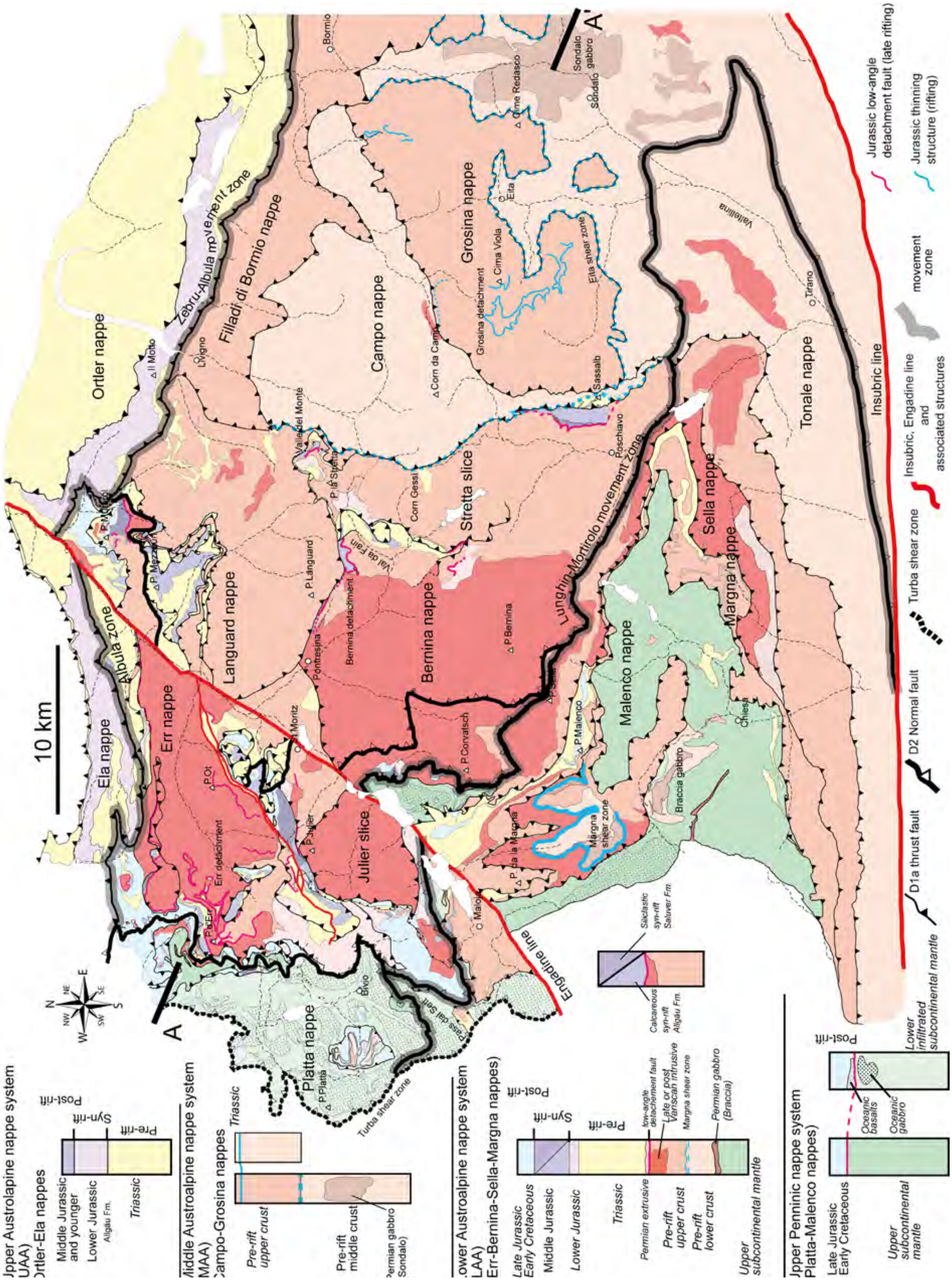


Fig. II. 3: Geological map of the Austroalpine and Upper Penninic nappes in SE-Switzerland and N-Italy, between Valtelina, Albula, and Julier valleys. Map compiled after Cornelius (1932, 1935, 1950); Staub (1946); Bonsignore et al. 1969; Montrasio et al. 1969; Beath et al. (1987); Liniger (1992); Spillmann (1993, 2005); Froitzheim et al. (1994); Manatschal (1995); Meier (2003); Trommsdorff et al. (2005) Peters (2005, 2007) and own observations.

(1) Upper Austroalpine nappe system (UAA) (Ortler-Ela nappes) derived from the former proximal margin and characterized by pre-rift upper crust, fault bounded rift basins and thick pre-rift sediments associated with calcareous early Jurassic (pre-Pliensbachian) syn-rift sediments (Allgäu Formation (Fm.)) (for a review see: Eberli 1988; Bernoulli et al. 1990)

(2) Middle Austroalpine nappe system (MAA) (Campo/Tonale-Grosina/Languard nappes) derived from the former necking zone and characterized by pre-rift middle and upper crust, with discontinuous and/or lacking pre- and syn-rift sediments

(3) Lower Austroalpine nappe system (LAA) (Err-Bernina-Margna and Sella nappes) derived from the former distal margin (Froitzheim and Eberli 1990; Spillmann 1993; Handy 1996; Hermann and Müntener 1996; Manatschal and Nievergelt 1997) and characterized by pre-rift upper and lower crust, the occurrence of low-angle detachment faults, discontinuous pre-rift and siliciclastic (Saluver Fm. e.g. Finger 1978) late Early Jurassic (post-Pliensbachian) to Middle Jurassic syn-rift sediments.

In order not to confuse the reader with too many names for local, Alpine tectonic nappes, we group the names of tectonic nappes used in this paper and show their distribution in Fig. 4.

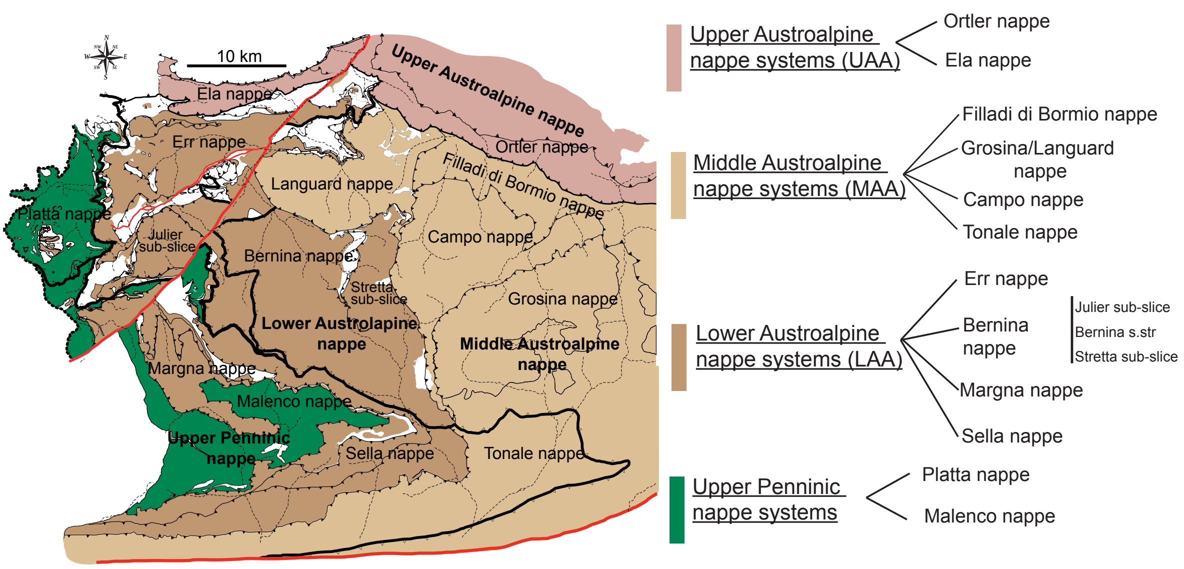


Fig. II. 4: Tectonic map of Austroalpine and Upper Penninic units in SE Switzerland and N-Italy showing the distribution, position and names of tectonic nappe systems and nappes used in this study. This classification is based on the work of Trumpy (1980).

3. Nature of basement rocks and sedimentary cover

The basement rocks of the Austroalpine nappe systems in SE-Switzerland and N-Italy record a complex polyphase magmatic and metamorphic evolution that is in detail difficult to unravel. However, the presence of post Variscan Permian and Jurassic intrusive and extrusive magmatic rocks as well as Mesozoic pre-, syn- and post-rift sediments enables to distinguish between Variscan, post-Variscan (Permian and Jurassic) and Alpine tectonic and metamorphic events. Additional constraints come from thermo-chronology work that provides information on the age of crystallization and cooling of the various basement rocks forming the Austroalpine nappe stack.

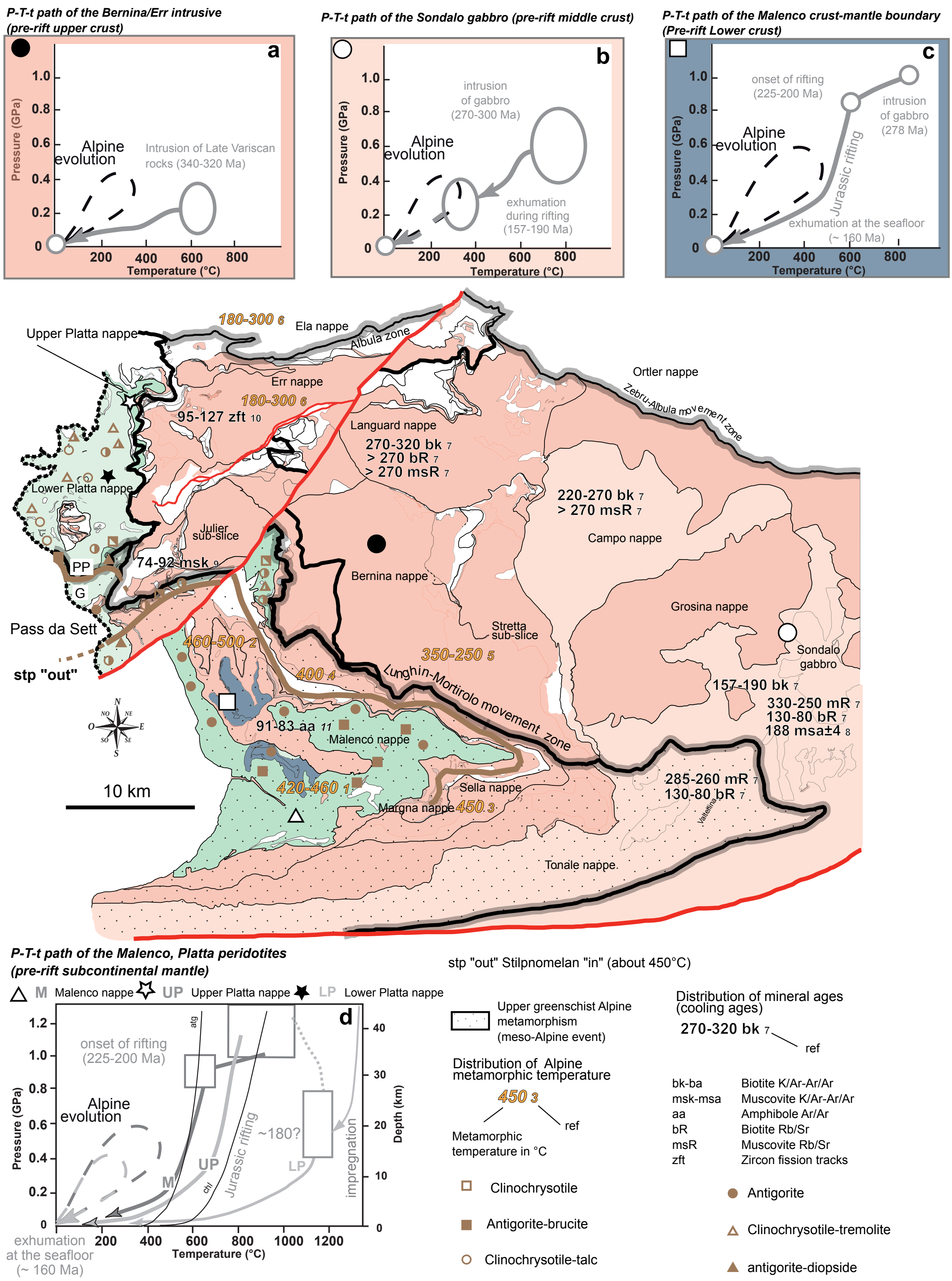
3. 1. Pre-Permian events

The pre-Permian basement consists of Variscan or older poly-metamorphic rocks composed of various lithologies that show in most cases an amphibolite-facies metamorphic overprint. These rocks experienced typically temperatures between 550 and 700 °C during the Variscan orogeny (e.g. 650-700°C for the Err nappe (Halmes 1991), 550-700°C for the Campo nappe (Bucher and Frey 1994)). The age of these parageneses is poorly constrained. Dating of similar mineral parageneses from the Oetztal basement (Upper Austroalpine) and the Southern Alps indicates a peak of metamorphism around 320-340 Ma (Thöni 1999; Boriani and Villa 1997), which is consistent with a Variscan event. Based on cross cutting relationships between felsic intrusive rocks emplaced in upper crustal levels during the Variscan orogeny and basement structures, it appears that many structures are Variscan or even older (Von Quadt et al. 1994). Most of the felsic intrusives show a calcalkaline trend and predate the emplacement of alkaline syenite-alkaligranitic intrusives (Rageth 1984; Spillmann and Büchi 1993). Dating of these rocks indicate an emplacement age of about 338-324 Ma (von Quadt et al. 1994). The P-T conditions determined from these rocks indicate a very shallow depth of intrusion (< 3km) (Peters 2005, 2007). These Variscan intrusive rocks are not affected by a metamorphic or deformation event, which contrasts with most crystalline basement rocks found in the Alps. This is well indicated by the preservation of the contacts between late Variscan intrusive and country rocks with no evidence of post-magmatic deformation (Staub 1916; Rageth 1984; Spillmann and Büchi 1993). Most of these pre-Permian rocks forming the basement of the Alpine nappes were in a quite shallow crustal position already at the end of the Variscan orogeny.

3. 2. Permian magmatic and metamorphic overprint

The post-Variscan Permian event is a key for the understanding of the lithospheric evolution of the Alpine domain in general and the Austroalpine nappes in particular. The geodynamic interpretation of the Permian event remains little constrained and its importance was probably underestimated. More recent studies by Schuster and Stüwe (2008) and Froitzheim et al. (2008), show, however, that this event has major implications for the subsequent Jurassic rifting and Alpine overprint. This is mainly due to the fact that a Permian overprint can be found at all crustal levels as well as in the underlying subcontinental mantle.

In the mantle, the Permian event is recorded by the depletion and resetting of the Sm/Nd isotopic system (Rampone et al. 1996, 1998). This event is commonly interpreted as being linked with a thermal re-equilibration of the lithosphere. The magmatic rocks formed during this event can be found in all crustal levels. They discordantly cross cut the Variscan and pre-Variscan fabrics, which enable us to distinguish between pre-Permian (Variscan or older), Permian fabrics and younger metamorphic events. At the surface, the Permian event is documented by volcanic and volcanoclastic sediments that infill Permian grabens (Bertotti et al. 1993, e.g. Ruina Fromation of Furrer et al. (1985)). Froitzheim et al. (2008) proposed that these grabens are related to low-angle detachment systems, suggesting crustal extension. Intrusions of felsic bodies in shallow crustal levels are documented from various places in the Alps (e.g Schaltegger and Brack 2007). In our study area, such events are marked by the extrusion of alkaline rhyolitic rocks dated with U/Pb on Zircon at 295-288 Ma (von Quadt et al. 1994) and associated with graben formation. The partial melting of the asthenosphere produced the intrusion of mafic gabboric bodies in lower (Malenco/Margna and Ivrea) (Handy et al. 1999 and references therein; Müntener et al. 2000 and Hermann et al. 2001) and middle (Sondalo, Mont Collon) (Tribuzio et al. 1999; Braga et al. 2001, 2003; Montjoie et al. 2005, 2007) crustal levels, in rare examples also in upper crustal levels (Bocca di Tenda Gabbros) (Tribuzio et al. 2009). These emplacements were associated with a contact metamorphic overprint. In our area, petrological studies from the Malenco/Margna nappes indicated that these rocks underwent isobaric cooling during the Triassic (Mïntener et al. 2000) (Fig. 5). Exhumation of these rocks initiated in Late Triassic to Early Jurassic time (Müntener et al. 2000; Villa et al. 2000). This is well shown in the Malenco/Margna nappes, where Early Permian gabbroic rocks intruded the crust-mantle transition at 270+6-4 Ma (U/Pd on Zircon; Hansmann et al, 1996) and were cooled (270-200 Ma) and exhumed to the seafloor during Jurassic time (200 Ma-130 Ma) (Villa et al. 2000). Permian gabbros (e.g. Sondalo Gabbro in Fig. 3) are also found in the Campo basement


P-T-t path of the Bernina/Err intrusive (pre-rift upper crust)
a
Pressure (GPa)
Alpine evolution
Intrusion of Late Variscan rocks (340-320 Ma)
Temperature (°C)
P-T-t path of the Sondalo gabbro (pre-rift middle crust)
b
intrusion of gabbro (270-300 Ma)
exhumation during rifting (157-190 Ma)
P-T-t path of the Malenco crust-mantle boundary (Pre-rift Lower crust)
c
onset of rifting (225-200 Ma)
intrusion of gabbro (278 Ma)
Jurassic rifting
exhumation at the seafloor (~ 160 Ma)
180-300 6
Ela nappe
Albula zone
Upper Platta nappe
Err nappe
Ortler nappe
Zebru-Albula movement zone
95-127 zft 10
Languard nappe
270-320 bk 7
> 270 bR 7
> 270 msR 7
Lower Platta nappe
220-270 bk 7
> 270 msR 7
Campo nappe
Julier sub-slice
74-92 msk 9
Bernina nappe
PP
G
Grosina nappe
Pass da Sett
Stretta sub-slice
460-500 2
350-250 5
Sondalo gabbro
stp "out"
400 4
Lunghin-Mortirolo movement zone
157-190 bk 7
330-250 mR 7
130-80 bR 7
188 msa±4 8
91-83 aa 11
Malenco nappe
10 km
420-460 1
Sella nappe
285-260 mR 7
130-80 bR 7
Margna nappe
450 3
Tonale nappe
P-T-t path of the Malenco, Platta peridotites (pre-rift subcontinental mantle)
M Malenco nappe
UP Upper Platta nappe
LP Lower Platta nappe
d
atg
chl
~180?
impregnation
Depth (km)
exhumation at the seafloor (~ 160 Ma)
stp "out" Stilpnomelan "in" (about 450°C)
Upper greenschist Alpine metamorphism (meso-Alpine event)
Distribution of Alpine metamorphic temperature
450 3
ref
Metamorphic temperature in °C
Distribution of mineral ages (cooling ages)
270-320 bk 7
ref
bk-ba Biotite K/Ar-Ar/Ar
msk-msa Muscovite K/Ar-Ar/Ar
aa Amphibole Ar/Ar
bR Biotite Rb/Sr
msR Muscovite Rb/Sr
zft Zircon fission tracks
Clinochrysotile
Antigorite-brucite
Clinochrysotile-talc
Antigorite
Clinochrysotile-tremolite
antigorite-diopside

(Del Moro and Notarpietro 1987; Tribuzio et al. 1999). However, based on the relations with the surrounding basement rocks as well as P-T estimates of the country rocks of the gabbros, Tribuzio et al. (1999) concluded that these rocks were intrusive in mid-crustal levels (0.6±0.2 GPa) between 300 to 270 Ma (Fig. 5). This intrusion was associated with pegmatites cross-cutting the former Variscan fabrics. All these different magmatic rocks were dated as Permian using different methods (for a review see Marotta and Spalla 2007). Because these rocks are often associated with granulites, P-T-t conditions estimated for these rocks enable us to use them as markers to define the position of the different rocks in the crust before Jurassic rifting (Fig 5). Based on these results, it can be shown that the basement rocks forming the Lower Austroalpine (LAA) and Middle Austroalpine nappe system (MAA) derived from pre-rift, i.e. Permian upper, middle and lower crustal levels (Fig 3). Since all basement rocks have normal stratigraphic contacts with Middle Jurassic and younger sediments, their exhumation must have occurred after Permian but before or during Middle Jurassic time. Cooling ages show that most of these rocks were exhumed during Middle Jurassic time (for discussion see below).

3. 3. Stratigraphic record (Fig. 6)

Pre-rift sediments

The stratigraphy of the Permian to Middle Jurassic sedimentary formations of the Austroalpine nappe systems in SE Switzerland have been extensively studied by the group of Rudolf Trumpy (Dössegger 1974, Furrer 1981; Eichenberger 1986, Naef, 1987, Finger 1978; Eberli 1988). The stratigraphic record starts with a volcano-sedimentary sequence of Permian age (Ruina Formation of Furrer et al. (1985))), which includes tuffs and ignimbrites (von Quadt et al. 1994). The overlying Permo-Triassic sediments record the transgressional development from conglomerates sand- and siltstones composed of crystalline components (Chazforà and Fuorn Formation of Dössegger (1974)) to a shallow marine thick carbonate plate-form. This carbonate plate-form can be subdivided in three sub-units: a lower Middle Triassic plate-form

Fig. II. 5: Map showing a compilation of the distribution of Alpine metamorphism and the isotopic ages in the Austroalpine and Upper Penninic nappe systems in SE Switzerland and N-Italy (data compiled from (1)Hermann and Müntener (1992); (2-3) Linger (1992); (4-5) Spillmann (1993); (6) Ferreiro Mählmann (1996); (7) Thöni (1981); (8) Meier (2003); (9) Handy et al. (1996); (10) Eggenberger (1990); (11) Villa et al. (2000). Serpentine paragenesis after (Trommsdorff and Evans 1974; Trommsdorff 1983). P-T-t paths are shown for (a) pre-rift upper crustal granitoids (Bernina and Err intrusives (Peters 2005, 2007)); (b) pre-rift middle crustal gabbros (Sondalo gabbro (Tribuzio et al. 1999; Bragga et al. 2001, 2003)); (c) pre-rift lower crustal rocks (Malenco crust – mantle boundary (Müntener et al. 2000; Villa et al. 2000)); (d) pre-rift subcontinental mantle (Malenco and Platta (Müntener et al. subm and references therein)). The data highlight the exhumation during Jurassic rifting of middle to lower crust as well as subcontinental. Note that based on the distribution of Alpine metamorphism and thermo chronological data two nappe edifices can be distinguished that are separated by the Lunghin – Mortirolo movement zone (see text).

(e.g. S-charl, Ducan, Vallatscha, Prosanto, Altein, Parai-Alba Formations of Eichenberger (1986); Franck 1986 and Naef 1987), an evaporite bearing sequence of Carnian age (e.g. Raible Formation of Frank 1986), and an upper massive dolomite plate-form of Norian age (e.g. Hauptdolomit Formation of Furrer (1981) or Dolomia principale of Bonsignore et al. 1969 and Berra 1995). The fauna as well as the sedimentary structures observed in these two carbonate plate-forms are characteristic for a shallow marine depositional environment with transitory subaerial exposure. Evidence for volcanic activity is indicated by intrusion of basaltic sill and deposition of ash layers dated in the lower plate-form at 240,91±0,26 Ma and 239,89±0,21 Ma (Furrer et al. 2008). The plate-form develops at its upper part into dark shales interbedded with lagoonal riff limestones and dolomites that were deposited in a protected lagoonal environment during Rhaetian time (e.g Kössen Formation of Furrer, 1981). The increase in subsidence observed in late Norian and Rhaetian time, as well as the occurrence of locale high-angle faults (Berra 1995) may represent the onset of rifting that leads to the opening of the Alpine Tetyhs ocean (Froitzheim and Eberli 1990). Overall, the Permo-Triassic sedimentary sequence represents, where complete, a more than 2 km thick pre-rift sequence (Bernoulli et al. 1990). However, this is only true for the UAA where the Permo-Triassic sequence is continuous and only slightly fragmented by Jurassic and Alpine deformation. In the MAA and LAA, the pre-rift sequence is highly disturbed and discontinues, which we interpret, as discussed later in this paper, as a result of Jurassic rifting.

<u>Syn rift sediments in the proximal margin (Upper Austroalpine domain)</u>

From Late Triassic onwards sedimentation in the various Austroalpine domains evolved in a very different way. In the UAA, rifting is linked to the formation of high-angle normal faults bounding major rift basins (e.g. Froitzheim 1988; Eberli 1988; Bernoulli et al. 1990). These basins were filled by breccias and calciturbides interbedded with hemipelagic marls and limestone belonging to the Allgäu Formation (Gümbel 1856; Jakobshagen 1965; Eberli 1988). The sediments filling the basin were mainly derived from the erosion of the Triassic carbonate plate-form. Eberli (1988) showed that the Allgäu Formation is divided in two main sequences, a lower and an upper sequence that are separated by middle Jurassic (Toarcian) manganese-bearing shales (Eberli, 1988). These shales represent an important correlation horizon, which has been dated in the Northern Calcareous Alps and in the Southern Alps as early Toarcian (Jakobshagen 1965; Jenkyns et al. 1985; Jenkyns and Clayton 1986). In contrast to the lower sequence, the age of the upper sequence is badly constrained and its transition into the first post-rift sediments is ill defined. Because all major rift-related faults in the UAA are sealed already by intra Allgäu sediments dated as Upper Sinemurian (Foitzheim and Eberli 1990), the upper sequence may be interpreted locally as post-tectonic. The upper sequence is overlain by the Radiolarian cherts, dated in the Alpine Tethys domain as Bathonian to Oxfordian (Baumgartner 1987; Bill et al. 2001). Thus Bathonian may be considered as a maximum age for the Allgäu Formation. The Radiolarian cherts are the first sediments overlying exhumed mantle and pillow basalts in the Upper Penninic nappe system. Therefore, on the scale of the margin, the Radiolarian cherts can be interpreted as the first post-rift sequence. It is important to note here that the age of the syn-rift sediments changes across the margin (see Froitzheim and Eberli 1990). Therefore the use of the term "syn-rift" is problematic (for further discussions see Péron-Pinvidic and Manatschal (2009)).

Syn to post-rift sediments in the distal margin (Lower and Middle Austroalpine domains)

The LAA and MAA show a distinct stratigraphic evolution starting from the late Triassic. Indeed in the Lower Austroalpine domain there is a lack of evidence for true extensional activity during early Jurassic time, as indicated by the occurrence of a carbonate platform, which shows neither significant aggradation nor evidence for fault bounded basins or lateral thickness variations. Thus, based on the stratigraphic record, it looks as if the Lower Austroalpine domain was only little extended during initial rifting (e.g. Eberli 1988). The transition from the Kössen Formation (latest Triassic) into the Lower Jurassic Agnelli Formation

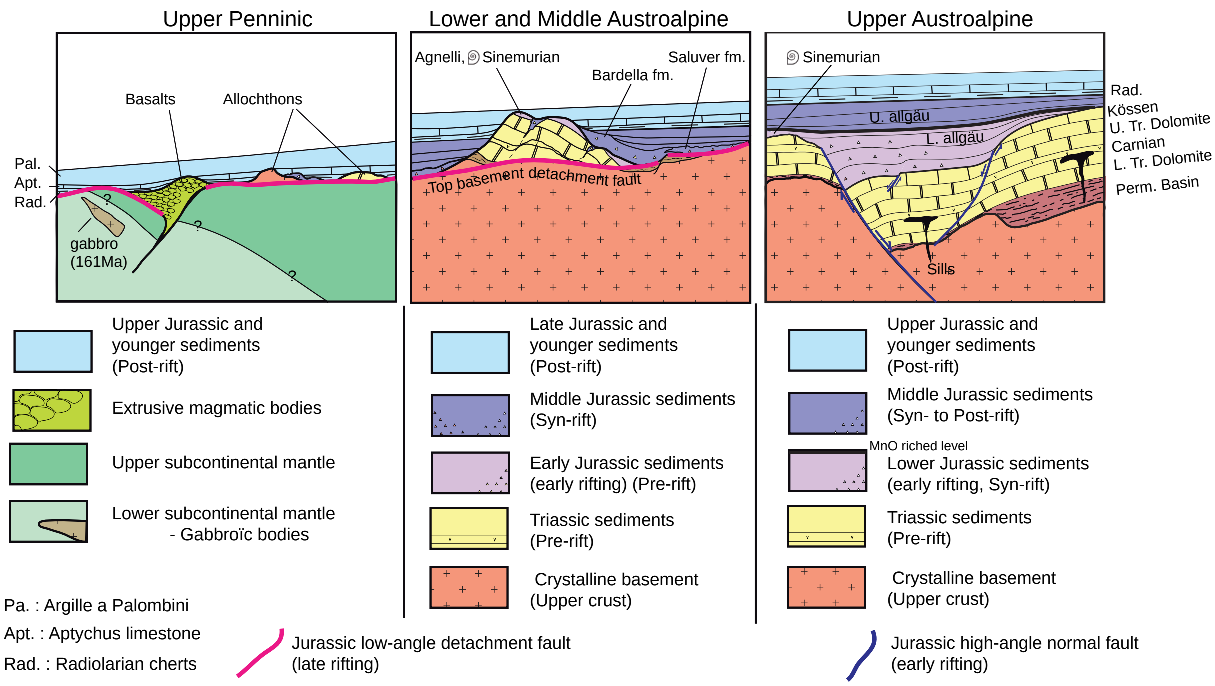


Fig. II. 6: Stratigraphic architecture of the Mesozoic sediments observed in the Austroalpine and Upper Penninic nappes in SE-Switzerland and N-Italy. Cartoons exemplify the observed stratigraphic and tectonic relationships observed in the UAA, MAA, LAA and Upper Penninic nappes.

is made by dark thin bedded limestones passing into massive light siliceous limestones with layers and nodules of chert indicating deposition in a hemipelagic environment (Finger 1978). This massive, about 50 m thick limestone has been dated with ammonites (Paltechioceras sp.) giving a Sinumurian age (Finger 1978). The top of the Agnelli Formation is capped by a hard-ground containing ammonites, which give an early Pliensbachian age (Manatschal and Nievergelt 1997). This evolution contrasts with that of the Upper Austroalpine domain, where at the same time high-angle normal faults associated with the formation of important rift basins occurred. It is important to note that the Agnelli Formation is the time equivalent of the lower Allgäu sequence in the UAA.

The first syn-tectonic deposits found in the Lower Austroalpine consist of locally derived breccias, debris flows and turbidites (e.g. Alv/Bardella Formation of Schüpbach (1973) and Finger (1978)). They are associated with high-angle faults affecting and eroding the Lower Jurassic and Triassic plate-form (Froitzheim 1988). In the Bernina nappe these breccias are preserved in synforms (Mezzaun, Alv, Sassalb) (Fig. 3) and are interfingered with sediments belonging to the upper Allgäu Formation. The best example is exposed in the Val dal Fain – Piz Alv area, where sediments from the Allgäu Formation lie directly on the basement with an onlap of less than 30°, a situation that has never been described from the UAA. In the Err nappe, the Bardella Formation is laterally replaced by detritic, basement derived breccias, ranging from tectono-sedimentary to sedimentary breccias that develop upsection into fine-grained distal turbidites (e.g. Saluver Formation of Finger 1978). These clastic sediments overlie exhumed crustal rocks (see Fig. 11 of Manatschal and Nievergelt 1997). The occurrence of a local siliciclastic source in the distal margin associated with the exhumation of crustal rocks highlight the onset of exhumation of basement rocks along top-basement detachment faults. It is important to note that the detachment faults serve as a new sedimentary source but at the same time, they also created the space and form the substrate onto which the syn-tectonic sediments are deposited. Thus, the Saluver Formation marks the change from high- to low-angle detachment faulting (see Mohn et al. 2010). The maximum age of this formation is constrained by the Pliensbachian hardground of Agnelli Formation onto which the syn-rift sediments onlap (Manatschal 2004). The upper boundary is constrained by Radiolarian cherts, which are deposited between Bathonian to Oxfordian time (Baumgartner 1987; Bill et al. 2001). It gives us an interval of approximately 20 Ma for the deposition of the Saluver Formation. The Radiolarian cherts are the first post-rift sediments, in the sense that they are the fist sediments that overlie embryonic oceanic crust. They are not continuous throughout the study area and it can be observed that elsewhere basement and pre-rift sediments (e.g. extensional allochthons) are overlain directly by Calpionella limestone (often also referred to as Aptychus limestone) and Argille a Palombini (Weissert and Bernoulli 1985). This is compatible with the occurrence of reworking processes within the Radiolarian cherts due to gravity processes. This suggests that a rift related topography prevailed during the post-rift sedimentation. The Radiolarian cherts are overlain by micritic limestone with intercalations of shales (Aptychus limestone or Calpionella limestone) that are conformably overlain by the Argille a Palombini Formation made by dark siliceous shales and calcarenites alternating with dark grey limestones (Weissert and Bernoulli 1985). This formation has been dated by Dietrich (1969) who found planktonic foraminifera of Aptian to Albian age. In few localities in our study area we can find polygenic breccias associated with the Couche Rouge Formation dated as Cenomanian (Peters 2005).These sediments represent the youngest sediments observed in the study area.

4. Alpine metamorphic overprint (Fig 5)

Mesozoic sediments and ophiolites are the key lithologies to characterize the Alpine metamorphic overprint and to distinguish it from the pre-Mesozoic metamorphic history. The distribution of Alpine metamorphism shows major differences between the northern and southern part of the study area (e.g. Spillmann and Büchi 1993; Froitzheim et al. 1994; Handy et al. 1996). Previous authors (Trommsdorff 1983, Früh-Green et al. 1990) suggested a gradual increase of the metamorphic overprint from N to S. This interpretation was supported by the observation that the Alpine metamorphic overprint increases from diagenesis to anchizone in the north (Totalp and Silvretta nappes), to prehnite-pumpellyite facies in the northern Platta, Err, Bernina, Grosina/Languard and Campo nappes (Dunoyer de Segonzac and Bernoulli, 1976; Biehler 1990; Ferreiro Mählmann 1994, 1996) to greenschist to lower amphibolite facies in the southern Platta, Margna, Sella and Malenco nappes (Trommsdorff and Evans 1977; Hermann and Müntener 1992; Liniger 1992; Spillmann 1993; Bissig and Hermann 1999) (Fig. 5). However, a major metamorphic gap exists at the boundary between the northern Platta, Err, Bernina and Campo nappes in the hanging wall and the southern Platta, Margna, Sella, Tonale and Malenco nappes in the footwall. This gap can be mapped along a line that can be traced between Pass dal Sett to Valtelina (Figs. 3, 5). North of this line, the Alpine metamorphic overprint never exceeded 350°C (Dietrich 1969). Dietrich (1969) and Ferreiro Mählmann 1996 showed that the metamorphic overprint within this area is very heterogeneous and was probably overestimated in the past. In contrast, to the south of this line, the rocks are at upper greenschist to epidote-amphibolite facies conditions as highlighted by the absence of stilpnomelane (Liniger 1992). In the Malenco nappe, the Alpine pressure peak is inferred from Na-bearing amphiboles at about

0,5-0,6 GPa (Bissig and Hermann 1999), followed by peak temperatures around 400-450°C, determined by calcite-dolomite and garnet-amphibole thermometry (Trommsdorff and Evans 1977; Mellini et al. 1987; Hermann 1997) (Fig. 5). Similar metamorphic conditions are also described from the Margna nappe (Guntli and Liniger 1989; Bissig and Hermann 1999).

The structure that limits the less-metamorphosed hanging wall from the more penetratively metamorphosed and deformed footwall is a complex, polyphase deformation zone, here referred to as the "Lunghin – Mortirolio" movement zone (for further details see discussion below) (Fig. 5). The change in metamorphic grade across this zone is also shown by a change in the style of deformation from localized Alpine deformation in the hanging wall to distributed and penetrative deformation in the footwall.

Pre-Alpine exhumation history and Alpine thermal overprint recorded in the Austroalpine nappes (Fig. 5)

In the past, geochronological investigations focused mainly on the Late Variscan and Permian intrusive rocks (e.g Braccia gabbro (Hermann 1997; Müntener et al. 2000); Sondalo gabbro (Tribuzio et al. 1999), Val Ferrata and Brusiono granites, Serotini-Tremoncelli diorites and granodiorites (Del Moro et al. 1982, 1983; Boriani et al. 1985, Del Moro and Notarpietro, 1987) and pegmatites (Bachmann and Grauert 1981; Thöni 1981). Few thermo-chronological data exist at present from the Malenco, Margna, Sella and Tonale nappes. Nevertheless, the existing data show younger, mainly Late Cretaceous ages, indicating that the isotopic systems were probably reset during Alpine convergence. It is confirmed by the crystallization of new amphibole in this area (Villa et al. 2000), which is compatible with the observed upper greenschist to lower amphibolite metamorphic overprint. The thermo-chronological data, mainly $^{40}Ar/^{39}Ar$ on muscovite and K-Ar on biotite and muscovite obtained from the Err, Bernina, Grosina/Languard and Campo nappes are still preserving older, pre-Alpine cooling ages for istopic systems, which have closure temperatures ranging between 350°C and 450°C. This observation shows that isotopic systems have not been reset by Alpine overprint indicating that the metamorphic conditions in these nappes never reached the necessary temperatures. These conclusions support the distinction of two different thrust stacks that underwent a different Alpine metamorphic overprint and are separated by a major tectonic zone, the Lunghin – Mortirolo movement zone which we will discuss in a later part of this paper.

5. Deformation history

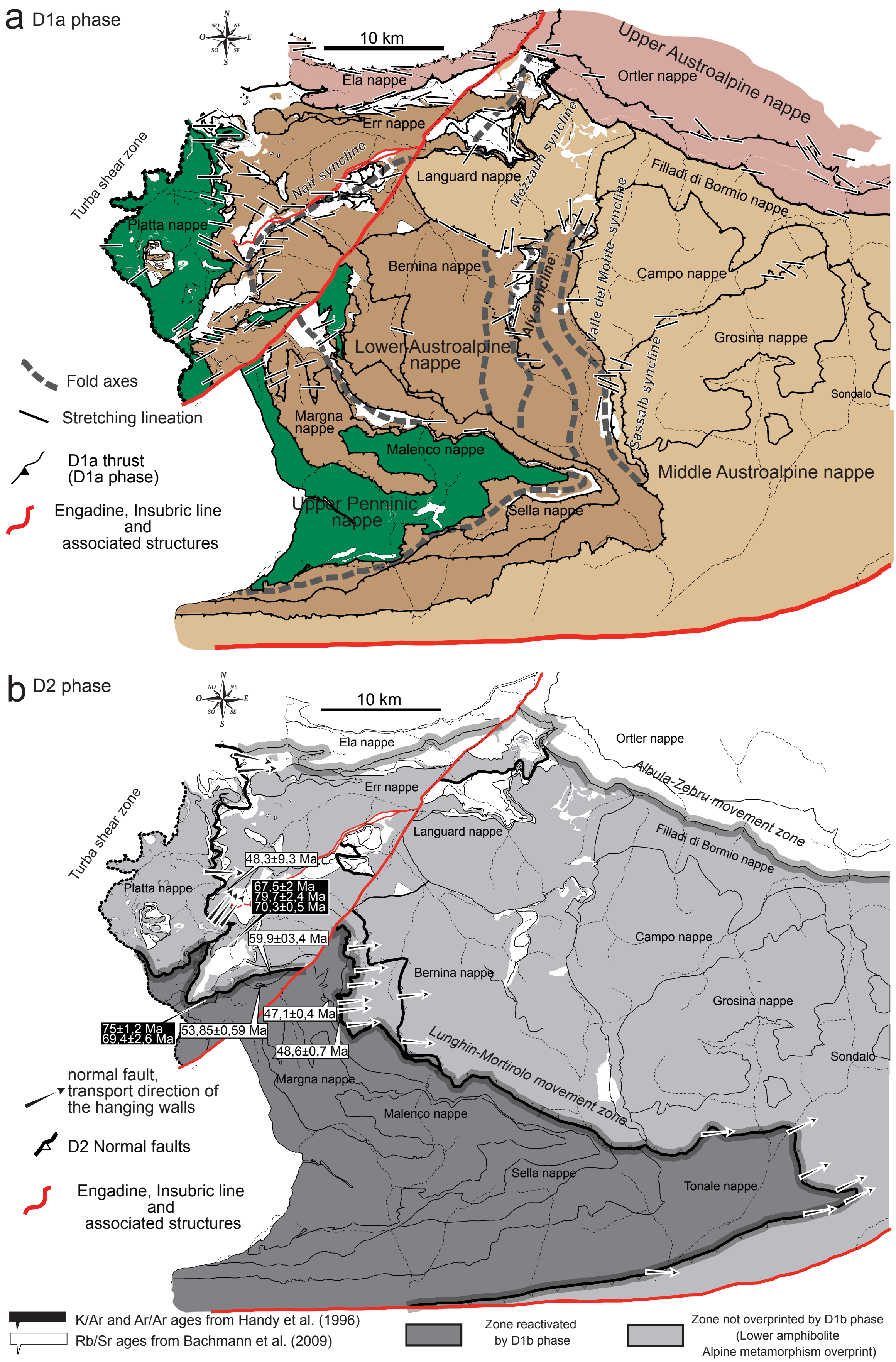
In this section, we review, based on own observations and the work of Liniger (1992); Spillmann (1993); Froitzheim et al. (1994) and Handy et al. (1996), the deformation history of the LAA and MAA. We will particularly focus on the Late Cretaceous and Tertiary evolution of these nappes (Fig. 1).

5. 1. D1 deformation phase

Previous authors (Liniger 1992; Spillmann 1993; Froitzheim et al. 1994) grouped all D1 structures related to W to NW directed nappe stacking within one and the same deformation phase (e.g. D1, the Trupchun phase of Froitzheim et al. 1994). Later, Handy 1996 recognized that D1 was a composite event involving an early stage of brittle thrusting followed by a later stage of ductile folding in the southern part of the Platta and Err nappe, at the boundary with the Margna nappe. However, mapping of the entire area containing the Platta nappe shows that south of a zone between Pass Lunghin and Mortirolo (Figs. 3, 7) the Platta, Malenco, Margna, Sella, and Tonale nappes show a penetrative ductile deformation linked with pressure-dominated metamorphism at upper greenschist- to lower amphibolite-facies conditions. This evolution contrasts with the lower Alpine deformation and metamorphic (prehnite-pumpellyite- to lower greenschist-facies) overprint reported from the northern Platta, Err, Bernina, Campo, Grosina, and Languard nappes. We argue that D1 structures are not all of the same age and did not form under the same P-T conditions. Thus, we distinguish between D1a and D1b structures.

5. 2. D1a deformation phase (Fig 7)

The first Alpine deformation event in the Austroalpine nappe systems, here referred to as the D1a event (see also Trupchun phase of Frotzheim et al. 1994; Handy et al. 1993, Handy 1996), is responsible for the main west to northwest directed stacking of the tectonic nappes reactivating the former proximal and distal Adriatic rifted margin. While the Upper Austroalpine Ortler and Ela nappes comprise mainly Triassic to Jurassic cover units derived from the former proximal passive margin, the LAA and MAA are more complex and consist of basement and reduced sedimentary sequences derived from the former distal margin (Fig. 2). Stretching lineations and associated sense of shear along the D1 fault plans indicate a top-to-the-west to -northwest directed transport direction (Froitzheim et al. 1994) (Fig. 7). The P-T conditions during this stacking event are poorly constrained. Handy et al. (1993) and Ferreiro Mählmann (1996) suggested a maximum temperature of about 300°C based on the brittle behavior of feldspar as well as the results of vitrinite reflectance and illite crystallinity values, whereas syn-


a D1a phase
10 km
Ela nappe
Err nappe
Upper Austroalpine nappe
Ortler nappe
Turba shear zone
Platta nappe
Nair syncline
Languard nappe
Mezzaun syncline
Filladi di Bormio nappe
Bernina nappe
Alv syncline
Valle del Monte syncline
Campo nappe
Grosina nappe
Lower Austroalpine nappe
Sassalb syncline
Sondalo
Margna nappe
Malenco nappe
Middle Austroalpine nappe
Upper Penninic nappe
Sella nappe
Fold axes
Stretching lineation
D1a thrust (D1a phase)
Engadine, Insubric line and associated structures
b D2 phase
10 km
Ela nappe
Ortler nappe
Err nappe
Albula-Zebru movement zone
Turba shear zone
Languard nappe
Filladi di Bormio nappe
48,3±9,3 Ma
Platta nappe
67,5±2 Ma
79,7±2,4 Ma
70,3±0,5 Ma
Campo nappe
59,9±03,4 Ma
Bernina nappe
Grosina nappe
47,1±0,4 Ma
75±1,2 Ma
69,4±2,6 Ma
53,85±0,59 Ma
48,6±0,7 Ma
Sondalo
Lunghin-Mortirolo movement zone
Margna nappe
normal fault, transport direction of the hanging walls
Malenco nappe
D2 Normal faults
Sella nappe
Tonale nappe
Engadine, Insubric line and associated structures
K/Ar and Ar/Ar ages from Handy et al. (1996)
Rb/Sr ages from Bachmann et al. (2009)
Zone reactivated by D1b phase
Zone not overprinted by D1b phase (Lower amphibolite Alpine metamorphism overprint)

kinematic growth of chlorite and white mica associated with the dynamic recrystallization of quartz indicate a minimum temperature of 250°C (Handy et al. 1996).

This deformation is localized along thrust faults in the footwall and associated with folding that is best expressed in the post-rift sediments. The folds range from the mm to the km scale and are associated with an axial plane foliation which is well developed in the limestones and shales of the syn- and post-rift formations (Handy et al. 1993). The foliation is marked by neoformation of illite and chlorite and opaque minerals (Manatschal and Nievergelt 1997). Growth of biotite is never observed. The folds are generally isoclinal with northeast to north trending axes (Froitzheim et al, 1994), however, in high strain zones as well as in areas with a strong inheritance, D1a fold axes are variably oriented (Handy 1996) and often occur as sheath folds (Manatschal and Nievergelt, 1997). The age of the D1a phase of deformation is constrained by both stratigraphic and radiometric ages. The youngest sediments affected by D1a structures belong to the Couches Rouge Formation, which is Cenomanian to lower Turoninan in age (Caron et al, 1982, Rösli, 1927, 1946).Thus a stratigraphic age for the D1a event is Turonian. Radiometric ages on D1 fault rocks from the Schlinig thrust (base of the Upper Austroalpine Otztal nappe) yield 100 Ma ages (Thöni 1986). In the LAA, Handy et al. (1996) obtained crystallization ages for D1a structures between 76±2,3 and 89,1±2,7 Ma based on the K-Ar method on white micas. The younger ages obtained from the LAA in respect to the UAA can be explained with the migration of the nappes from east to west (see Froitzheim et al. 1994 and references therein). Indeed the D1a deformation event is not related to the convergence of Adria and Europe, but associated with the Later Cretaceous closure of the Meliata-Vardar domain. The Austraolpine nappes studied in this paper were emplaced in a more external part at the D1a Orogen (Fig. 1). The general style of deformation and metamorphic overprint associated with the D1a event observed in the Austroalpine nappe systems in SE Switzerland show some similarities with fold and thrust belts observed in external parts of compessional systems. The fact that basement rocks are involved in the deformation in the Platta, Err and Bernina nappes can be explained, as discussed below, with the strong pre-Alpine inheritance.

Fig. II. 7: Structural map compiling the main structures related to (a) D1a and (b) D2 deformation phases. For D1a phase data are compiled from own observations and (Liniger and Guntli (1988); Spillmann 1993; Froitzheim et al. 1994 and references therein; Manatschal (1995) Handy et al. (1993, 1996)). For the D2 phase data are from Spillmann (1993); Froitzheim et al. (1994); Manatschal (1995) Meier (2003) (see text).

5. 3. D1b deformation phase

D1b structures are mainly observed in the southern part of the study area (e.g. southern Platta, Margna, Sella, Malenco and Tonale nappes). In these units D1b deformation overprints structures related to the emplacement of the nappes (D1a) as indicated by isoclinally folded nappe structures (Hermann and Müntener 1992; Handy 1996). The D1b event results in a strong and pervasive structural overprint that occurs under pressure-dominated conditions at 0.5-0.6 GPa (Bissig and Hermann 1999) dated by Villa et al. (2000) at 91-83 Ma, which is different from the deformation observed in the northern part. These results are in agreement with previous work of Handy et al. 1996 who documented a similar evolution based on Si content in white micas. We therefore suggest that Villa et al. 2000 dated the D1b phase related to onset of subduction in the Alpine Tethys domain. The 83-91Ma ages fall in the time range of Campanian to Santonian which corresponds with the time when subduction initiated in the Alpine Tethys.

Despite the same kinematics, the D1a and D1b structures form under different metamorphic conditions and cannot be explained with a single tectonic phase. Froitzheim et al. (1996) showed that active deformation during the D1 phase shifted from the eastern to the western Adria margin presumably leading to the onset of subduction in the Alpine Tethys domain within this time range. The transition from the D1a to the D1b system, which, occurred at around 84 Ma (Santonian/Campanian), corresponds in a shift of the subduction zone from the Vardar/Meliata to the Alpine Tethys domain. D1a structures cannot be interpreted as being related to the convergence of Adria and Europe. We agree with previous suggestions (Froitzheim et al. 1996, Schmid et al. 2008) that D1a nappe stacking is probably related to the final closure of the Meliata oceanic domain and therefore had to occur in an external position of this Late Cretaceous orogen (Fig. 1). This interpretation is coherent with (1) the occurrence of high-pressure metamorphism in the Upper Austroalpine basement nappes (2) The westward migration of the D1a across the entire Austroalpine realm. In contrast, the later, pressure-dominated D1b deformation localized south of the Lunghin-Mortirolo movement zone, may be related to the onset of subduction in the Alpine Tethys domain. The study area changed from a lower-plate to an upper plate position relative to the active subduction zone (Fig. 1) (Froitzheim et al. 1996).

5. 4. D2 deformation phase (Fig. 7)

The D2 phase (e.g. Ducan-Ela phase of Froitzheim et al. 1994) was interpreted as an extensional event. Observations indicating the occurrence of normal faults were first reported from the Silvretta nappe by Heim (1922); Eugster (1923) and Eichenberger (1986) (e.g. Ducan fault). Froitzheim et al. (1992, 1994) were, however, the first to describe an extensional event in the study area and to propose that the normal faults were kinematically associated with recumbent folds that were interpreted as collapse folds. In the study area D2 fault systems can be mapped based on their overprinting relationship with D1 structures. The observed normal faults cut through the D1a nappe stack and sole in the Lunghin–Mortirolo movement zone, indicating that this zone had also to be active during D2 extension. The most prominent D2 normal fault in the study area can be mapped from the Ela pass, through Val d'Err and the Julier Valley to Pass da Sett where it joins the Lunghin-Mortirolo movement zone. Another normal fault can be mapped at the base of Piz Corvatsch and Piz Bernina joining the Lunghin-Mortirolo movement zone in Val Malenco. All these faults show a top-to-the east sense of shear (Fig. 7b).

D2 structures have been dated by many authors. Handy et al. (1996) obtained ages between 67 and 80 Ma based on the K-Ar and $^{40}Ar/^{39}Ar$ methods respectively, on white micas and Na-amphibole derived from the Err and Platta nappes. More recently Villa et al 2000 interpreted 67-73 Ma old $^{40}Ar/^{39}Ar$ ages on amphibole from the Malenco nappe, based on compositional variations of the amphiboles indicating recrystalization under decreasing pressure, as possibly related to the D2 event. These ages may coincide with a thermal peak (Handy et al, 1996 and Villa et al, 2000), which occurred at around 75-67 Ma in the southern Platta, Margna, Sella and Malenco nappes. However, in a recent work, Bachmann et al. (2009) reported Rb-Sr ages on newly crystallized white mica from mylonitized meta-volcanics derived from a D2 shear zone at the base of the Err nappe that is folded by a major D3 fold in the Julier Valley between Bivio and Pass dal Julia. The authors interpreted their ages of 47,1±0,4 to 59,9±3,4 Ma, according to Müller et al. (1999), as related to the re-equilibration between white mica and coexisting phases during mylonitization at temperatures of about 350°C. Thus, the new data suggest that the D2 phase may be as young as 47Ma. D2 normal faulting could therefore be active in the Austroalpine nappe systems probably between 75 and 47 Ma, which corresponds with the time range during subduction in the Alpine Tethys domain. Rather than orogenic collapse, the D2 deformation phase could be linked to strain accommodation in the hanging wall of an active Alpine subduction.

5. 5. D3 deformation phase (Fig. 8)

The D3 deformation phase (e.g. Blaisun phase of Froitzheim et al. 1994) consists in the study area of large scale upright, open folds. Fold axes are generally sub-horizontal and strike constantly east-west, parallel to the mountain belt. In our study area the facing direction as well as wave length and amplitudes of the D3 folds varied from north-directed in the N, to south-directed in the S. Towards the south, close to the Insubric line, the wave length of folding decreases from the kilometer to the hundred meter scale (south of Pass dal Sett, Manatschal and Nievergelt 1997). In Val Malenco, the axial planes are documented by crystallization of new minerals indicating upper greenschist-facies condition (Hermann and Müntener 1992). However going further to the north, in Val d'Err, wave length increases and the axial plan of the D3 folds become vertical or dipping to the south (Handy et al. 1993; Handy 1996; Manatschal and Nievergelt 1997). The first major north-vergent D3-fold is the Ela fold that was interpreted as a D1 structure, overprinted by all subsequent deformation phases. In the LAA and MAA, south of the Engadin valley, D3 folds form kilometer scale synclinal and anticlinal structures with a weakly developed axial planar solution cleavage. Locally this deformation is documented by thrusting with north-south stretching lineations. This is particularly well observed in the Samedan zone in the Err nappe, were D3 faults with displacements of few 100 m offset D1 thrust faults (Handy et al. 1996).

The D3 phase forms the youngest pervasive structures in the LAA, MAA and Upper Penninic nappe system in the study area. This phase corresponds to the N-directed thrusting of the D1a nappe stack over the Middle and Lower Penninic nappe systems (Froitzheim et al. 1994). Based on the age of the youngest sediments found in the footwall of D3-thrusts a maximum age of middle Eocene (Ziegler 1956; Eiermann 1988) can be proposed. Based on the age, as well as on the change in polarity from north (pro)-to south (retro)-directed thrusting, this phase has been related to the Late Eocene continental collision between Adria and Europe. Due to the position of the study area in the hanging wall of the subduction zone (e.g orogenic lid of Laubscher 1983) and in the neutral zone between north and south vergent thrust faults in the Tertiary orogenic prism, the LAA, MAA and Upper Penninic nappe system in Central and SE Grisons remain relatively undisturbed and were weakly overprinted during the Eocene Alpine collision (see NFP 20 E cross sections of Schmid et al. 1996).

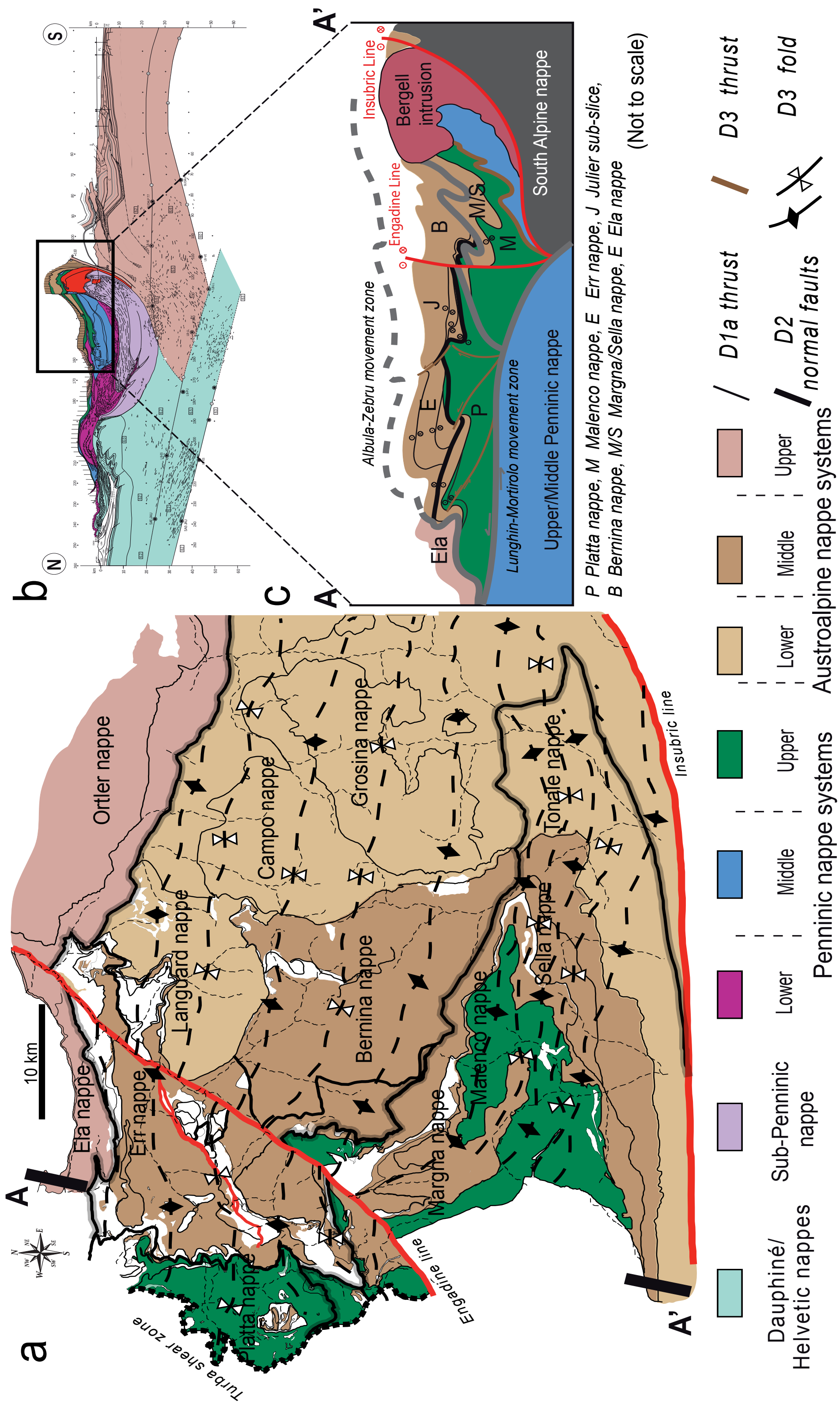


Fig. II. 8: (a) Structural map compiling the main structures related to D3 phase (data are from Manatschal 1995; Trommsdorff et al (2005) and own observations). (b) N-S oriented section across the Austroalpine and Penninic nappe systems in SE Switzerland and N-Italy (from Schmid et al. 1996) and (c) a section shows the change in polarity of the D3 structures from N to S vergent and their relation to the large scale orogenic structure. For the reconstructions shown in following Fig. 13 and 14 it is important to note that the studied zone studied is within the "neutral" zone of the compressional orogen, i.e. between the pro- and retro-vergent thrust units.

5. 6. Post D3 deformation (Turba shear zone and Engadin fault system)

The D3 structures are truncated in the western termination of the study area by a major east dipping, low-angle extensional shear zone referred to as the Turba mylonite zone (Nievergelt et al. 1996). This structure juxtaposes the D1a thrust-stack in the hanging wall against structures belonging to the Upper, Middle and Lower Penninic nappe systems in the footwall (Nievergelt et al. 1996). The Turba mylonite zone is truncated in the Val Bregaglia in the SW by the granodiorite of the Bergell intrusion, dated by von Blanckenburg (1992) at 30 Ma. This relationship gives a minimum age for the Turba mylonitic zone of 30 Ma. A maximum age is given by the fact that the Turba mylonitic zone cut and overprint D3 folds and overprints the Arblatsch flysch dated at 45-30 Ma. More recently, Pleuger et al. (2003) proposed that the time range for the activity of the Turba mylonitic zone is ranging between 35-30 Ma based on structural analysis on Adula, Tambo nappes and Misox zone located in hanging wall of the shear zone. However the geodynamic context of this shear zone remains poorly understood. Nievergelt et al. (1996) interpret the Turba mylonite zone to be linked with exhumation of the Lepontine dome in the Central Alps. In the study area, other major extensional structures belonging to the Turba phase are not observed.

The last major event affecting the study area is related to movements along the Engadine fault system, which is a branch of the Periadriatic system (Schmid et al. 1989; Schmid and Froitzheim 1993) (Fig. 3). This fault system affects the Austroalpine and Penninic nappe systems as well as the Bergell intrusion, which indicates that it must post-date the Bergell intrusion dated at 30Ma (von Blanckenburg 1992). The Engadine fault system shows a complex sinistral strike slip movement, with along strike changes from a vertical down-thrusting of the southern part in the SW to a down-thrusting of the northern block in the NW. Branches of the Engadine fault can also be observed to the north and south of the Engadine Valley. Schmid et al. (1989) and Schmid and Froitzheim (1993) showed that the offset along the main Engadin fault is between 2 and 3 km. This late fault system has only little impact on the former late Cretaceous nappe stack apart from the apparently offsetting related tectonic nappes across the fault, which resulted in some debates. Previous work by Schmid and Froitzheim (1993) suggested that the Ela nappe represents the continuation of the Bernina nappe north of the Engadin line. The Ela nappe was interpreted as part of the LAA meaning that it derived from the distal Adriatic margin. However, stratigraphic considerations (Eberli 1988) as well as comparisons of the syn-rift sediments and post-rift evolution do not support this interpretation. Indeed, the Ortler and Ela nappes preserve a continuous Mesozoic cover east and west of the Engadine line that is thrust over LAA and MAA along the same contact. Both the Ela and Ortler nappes have the same Mesozoic rift-related structure (Eberli 1988) characteristic at proximal margins. In contrast the Bernina nappe shows rift structures characteristic of a distal margin (e.g. Bernina detachment in Val dal Fain). In addition, the Ela and Ortler nappes show two distinct megacycles of sedimentation of the Allgäu Formation, whereas in the Bernina nappe only the uppermost cycle can be recognized (Fig. 16 of Eberli 1988). We thus conclude that the Ela nappe belongs to the UAA.

5. 7. The Lunghin-Mortirolo and Albula-Zebru movement zones

One important observation is that Alpine deformation reactivates and overprints inherited structures related to the Jurassic necking zone, resulting in complex polyphase deformation, here referred to as movement zones. In the study area two such movement zones can be defined: the Lunghin-Mortirolo and Ela-Zebru movement zones (for their location, see maps in Figs. 3 and 7). The fact that these two movement zones are truncated to the west by the Turba shear zone and to the south and southeast by the Insubric line indicates that these structures are pre-Oligocene in age.

The Lunghin-Mortirolo movement zone (Figs. 3, 7b) formed first as a top-to-west D1a thrust of Northern Platta, Err and Bernina nappes over the Southern Platta, Sella nappes. During D1b, this movement zone delimits a low-grade metamorphic D1a thrust stack in the hanging wall (northern Platta, Err, Bernina, Campo, Grosina, Languard) from an D1a thrust stack, overprinted by D1b deformation under pressure-dominated, upper greenschist- to epidote-amphibolite-facies conditions in the footwall (Figs. 3, 5, 7b, 8). The difference in metamorphic grade north and south of this zone is also shown by a change in deformational style from localized Alpine deformation in the hanging wall to distributed, penetrative deformation in the footwall. During Late Cretaceous extension (D2) several discrete, top-to-the-east normal faults flattened and reactivated the D1a Lunghin-Mortirolo thrust. Examples are the Corvatsch-Bernina boundary fault (Spillmann 1993), the base of the Julier fault (Froitzheim et al. (1994) and Handy et al. (1993)) and the Mortirolo shear zone (Meier 2003). As a consequence, the Lunghin-Mortirolo movement zone corresponds to a complex, several hundred meters- to kilometer-wide zone that assembled polyphase deformation at different conditions with different kinematics (see Figs. 5, 7a, b).

The Albula-Zebru movement zone can be mapped for a distance of over 30 km from Crap Ses in the Julier valley to Bormio in the Valtellina valley (Fig 3), juxtaposing UAA (Ela and Ortler) against LAA and MAA (Fig. 3). This structure was a top-to-the-west directed D1a thrust that was reactivated during the following deformation phases, which explains the complexity of this structure and the problem previous authors have faced in interpreting this zone as the result of one single deformation event (e.g. Zebru zone, Conti et al. 1994; Albula zone, Froitzheim et al. 1994). Because the Albula-Zebru movement zone truncates D3 fold structures in the Middle Austroalpine (Fig. 8c) but is truncated by the Turba mylonite zone and the Engadine line, the last movements along this zone had to have occured in Oligocene time. Like the Lunghin-Mortirolo movement zone, the Albula-Zebru shows a complex superposition of various deformational phases.

6. Cross sections across the Upper-Penninic and Lower and Middle Austroalpine nappe systems in Grisons

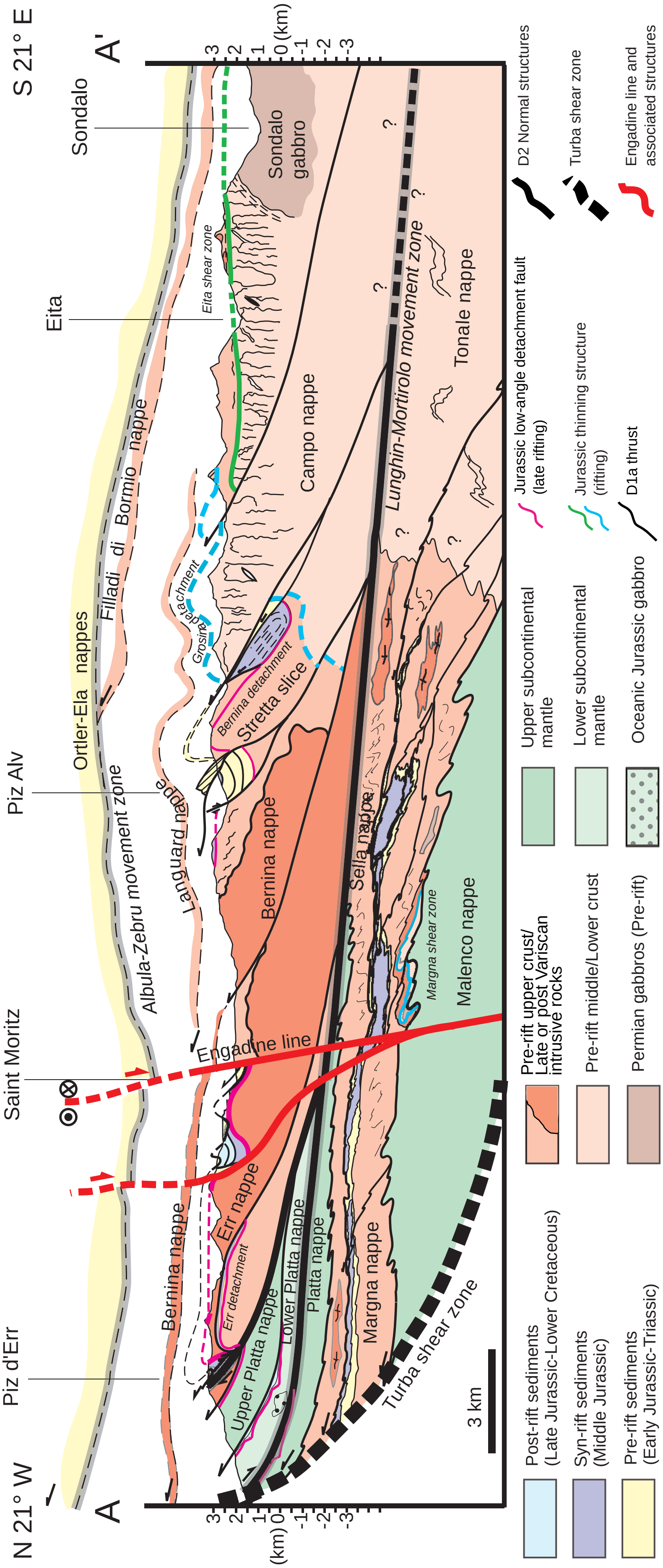
This study shows that the classical subdivision of the tectonic evolution of the Austroalpine nappe systems in SE Switzerland in deformation phases, as suggested by Froitzheim et al. (1994), underscores the importance of poly-phase reactivation of major deformation zones such as the Zebru-Albula and Lunghin-Mortirolio movement zones. These two structures can be mapped throughout the study area, separating three major tectonic "edifices" that are from top to bottom: (1) the UAA edifice, which preserves remnants of the former proximal margin (Ela and Ortler), (2) the LAA, MAA and Upper-Penninic nappe system edifice that preserves remnants of the former necking zone and the distal margin and ocean-continent transition (Grosina/ Languard, Campo, Bernina, Err and northern Platta nappes), and (3) The LAA and Upper-Penninic nappe system edifice that preserve remnants of the most distal margin and the OCT (southern Platta, Margna, Sella, and Malenco nappes).

In the cross section shown in Fig. 9, we focussed on a transect across the Platta, Err, Bernina, Grosina/Languard and Campo nappes affected mainly by the D1a event. Within this section, we focus mainly on three sectors: the Stretta - Campo - Grosina sector shown in Figs. 10, 11a,b; the Bernina - Stretta sector shown in Figs. 11c,d,e, 12, 13 and the Platta – Err sector that has been previously described by Manatschal and Nievergelt (1997). Indeed, the Alpine deformation structures observed in these sections are typical for a fold- and -thrust belt. They consist of folds associated with duplex structures and propagation thrusts. Mesozoic sediments and more particular syn- to post-rift sediments are generally preserved in the hinges of W-facing D1a synclines (Piz Alv, Piz Mezzaun, Piz Nair, Sassalb,Valle del Monte) (Figs 3, 6). These synclines are generally linked to drag folds related to D1 thrusts. Contacts between basement and sediments often form thrust faults.

6. 1. Alpine and pre-Alpine structures in the Campo – Grosina/Languard (Figs. 10, 11a,b)

A long standing problem in the MAA is the relationship between the Campo, Grosina and Languard nappes, and more particular the position of the former one. In the north, the Bernina nappe is overthrust by the Languard nappe along an D1a contact (Staub, 1946), while further south the Campo nappe is thrust over the Bernina nappe. Moreover the contact between Languard and Campo is not well exposed and therefore not well constrained. Based on field mapping we postulate that the Languard nappe was in a similar position to the Grosina nappe. This hypothesis implies that the Languard nappe overlies the Campo nappe and that the Languard and Grosina nappes are in an equivalent tectonic position. In this study we therefore referred to them, when we speak about the whole area as the Grosina/Languard nappe.

The Campo and Grosina nappes are essentially made of basement with almost no Mesozoic sediments. The contact between the Campo and Grosina nappes was first described by Staub (1946) and interpreted as a thrust fault due to the occurrence of thin slices of Mesozoic sediments (mainly dolomite) along their contact in Poschiavo valley. However the continuation of this thrust towards the south was subject of numerous different interpretations. In the classical interpretation, the thrust contact was interpreted to continue further eastwards into Italy. In this interpretation the Grosina nappe represents a klippe above the Campo nappe emplaced during the D1a phase. However this interpretation is not valid for the entire contact throughout the area. The northern portion records a top-the-W sense of shear indicated by stretching lineations and shear-sense indicators (shear bands, s-c fabrics) associated with similar mineral assemblage as adjacent D1a structures, whereas the eastern part (referred by Meier (2003) as the Eita shear zone) shows a different evolution (Figs. 9, 10). In the Grosina valley, the Eita shear zone located at the contact between the Campo and Grosina nappes contain stretching lineations defined by elongated K-feldspar porphyroclasts dipping gently to the NNW. Shear bands, mica fish, and asymmetric strain fringes around K-feldpars indicate a top-to-the-NNW sense of shear (Fig. 10). Meier (2003) estimated the temperature condition of the contact at around 400°C based on the stability of green-brown biotite, indicating that this contact formed at much higher metamorphic conditions. The significance of the eastern contact remains unclear but the


N 21° W
S 21° E
A
A'
Piz d'Err
Saint Moritz
Piz Alv
Eita
Sondalo
Ortler-Ela nappes
Filladi di Bormio nappe
Albula-Zebru movement zone
Languard nappe
Bernina nappe
Eita shear zone
Grosina detachment
Sondalo gabbro
Campo nappe
Tonale nappe
Lunghin-Mortirolo movement zone
Bernina detachment
Stretta slice
Sella nappe
Margna shear zone
Malenco nappe
Engadine line
Err nappe
Err detachment
Upper Platta nappe
Lower Platta nappe
Platta nappe
Margna nappe
Turba shear zone
3 km
(km)
Post-rift sediments (Late Jurassic-Lower Cretaceous)
Syn-rift sediments (Middle Jurassic)
Pre-rift sediments (Early Jurassic-Triassic)
Pre-rift upper crust/ Late or post Variscan intrusive rocks
Pre-rift middle/Lower crust
Permian gabbros (Pre-rift)
Upper subcontinental mantle
Lower subcontinental mantle
Oceanic Jurassic gabbro
Jurassic low-angle detachment fault (late rifting)
Jurassic thinning structure (rifting)
D1a thrust
D2 Normal structures
Turba shear zone
Engadine line and associated structures

kinematics as well as the metamorphic conditions are inconsistent with an D1a structure. In our map (Fig. 3), we trace the thrust of Grosina over Campo southwards, where it is cut by the Lunghin-Mortirolo movement zone (Figs. 3, 9, 10).

The Campo basement represents pre-rift middle crust as demonstrated by the intrusion of Permian gabbros (i.e. Sondalo gabbro) at 0.6±0.2 GPa and ~900°C (Braga et al. 2003 Tribuzio et al. 1999) (Figs. 3, 5 and 10). $^{40}Ar/^{39}Ar$ and K/Ar data by Thöni (1981) and Meier (2003) on biotite and muscovite from the Campo basement yield cooling ages of about 180 Ma. These data suggest that the Campo basement was in a mid-crustal depth before the onset of rifting. We interpret cooling ages of this domain to be linked with its exhumation during Jurassic rifting. These results are comparable with studies performed in the Malenco nappe, where Permian mafic rocks (i.e. Braccia gabbro) intruded pre-rift lower crust which was exhumed during Jurassic rifting (Fig. 5) (Müntener et al. 2000, Villa et al. 2000). Based on this structural, petrological and thermochronolgical data, we infer that the Eita shear zone does not represent an Alpine structure as previously suggested by Meier (2003) but was a Jurassic extensional shear zone active during rifting. The Grosina nappe has received little attention in the past and only few data are available from this nappe. The Grosina nappe is made by polymetamorphic basement intruded locally by felsic intrusive rocks similar to the rocks found in the Bernina nappe for which an upper crustal position during the Permian can be demonstrated. In the Grosina nappe, we mapped a major fault with a well-developed brittle deformation including characteristic black gouges (Fig. 10). These structures show many similarities to those described from the Jurassic Err detachment further to the north in the Err nappe (e.g. Fig. 5 of Manatschal 1999).

Within the Campo and Grosina basement, Cretaceous and Tertiary deformation is limited to very narrow tectonic horizons as shown in the Poschiavo valley at the boundary between the MAA and LAA. At Sassalb (Figs. 3, 10), Mesozoic sediments and basement rocks of the Bernina nappe are folded together with the Campo basement, forming a D1a synform related to a thrust (Figs. 3, 10 and 11a,b). The contact between the Campo and Bernina nappes at Sassalb is formed by a brittle fault containing very characteristic black gouges similar to those observed in the Grosina nappe. Detailed mapping of this contact juxtaposing Mesozoic sediments

Fig. II. 9: Reconstructed Alpine section across the Upper Penninic and Austroalpine nappes in SE-Switzerland and N-Italy. Note the different styles of Alpine deformation in the hanging wall and footwall of the Lunghin – Mortirolo movement zone (for localisation of the cross section see Fig. 3).

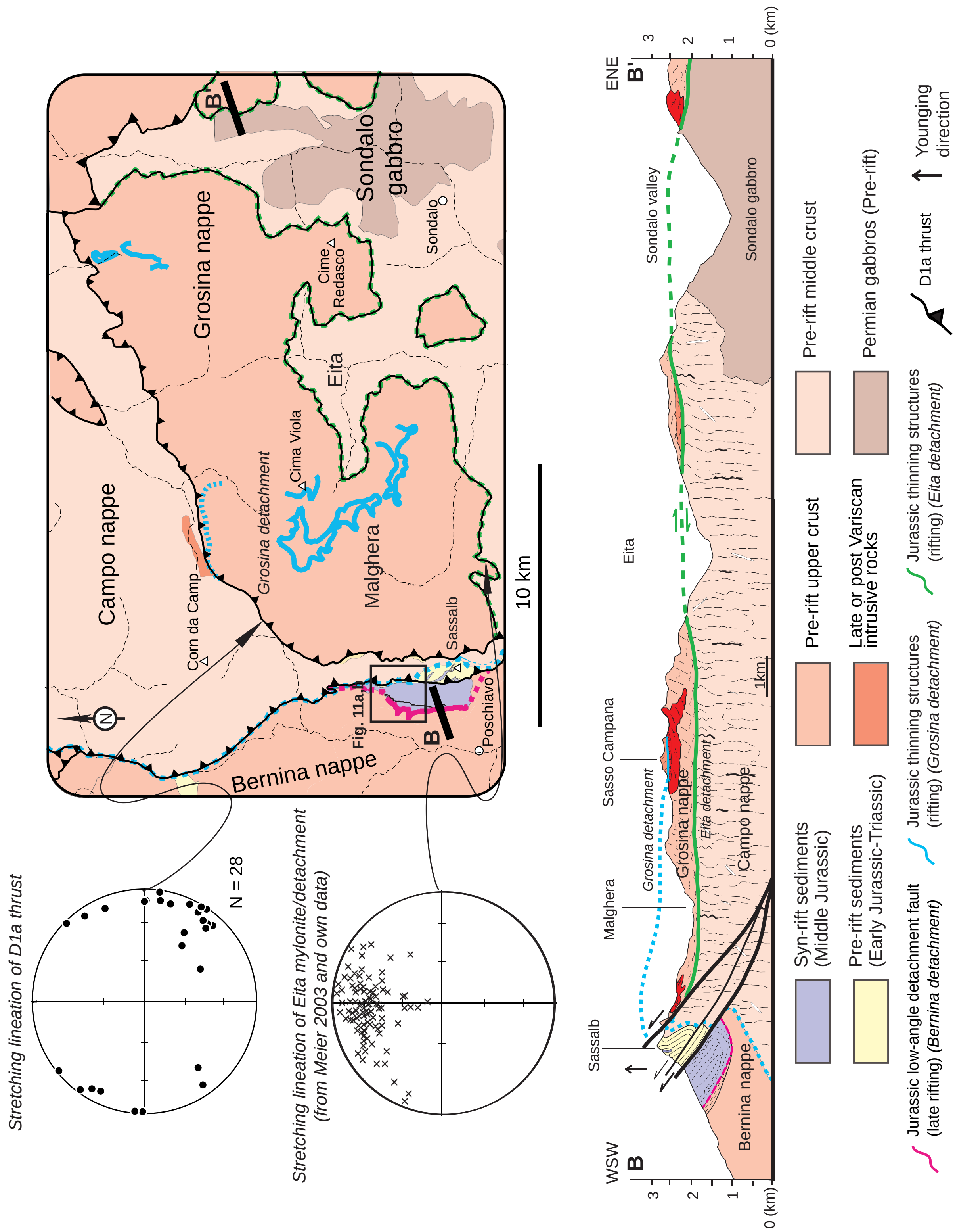


Fig. II. 10: Map and cross section across the Sassalbo – Sondalo area showing the relationships between the Grosina, Campo and Bernina nappes. The stereographic projection is in equal area-lower hemisphere representation.

against Permian mid-crustal basement rocks of the Campo basement shows that it is folded and truncated by D1a thrust faults (Figs. 10, 11a,b), suggesting that this contact must be pre-D1a in age. We interpret this fault as a pre-Alpine rift detachment referred as the Grosina detachment, which is supported by Jurassic cooling ages and the p-T-t path obtained from the Sondalo gabbro within the footwall of this fault (Fig. 5). The Grosina detachment can be correlated with the brittle fault mapped in the Grosina nappe (see Figs. 9, 10, 11a,b) separating Grosina-Campo in the footwall from the Bernina in the hanging wall.

Based on these observations, we conclude that the contacts between Campo and Grosina nappes (Eita shear zone) and between Bernina and Grosina-Campo nappes (Grosina detachment faults) represent rift-related structures. However, these structures are only locally preserved. In most parts of the study area, the rift-related structures were reactivated during Alpine convergence, indicating their importance during Alpine compression (Figs. 3, 10).

6. 2. Alpine and pre-Alpine structures in the Bernina nappe (Figs. 11, 12, 13)

The Bernina nappe s.l. can be subdivided in to three slices: the Julier, Bernina s.str, and Stretta slices (Fig. 3). The Julier slice is only exposed to the northwest of the Engadine Line and will not further be discussed in this paper. The displacement (~200 m) along the thrust splay separating the Bernina s.str. from the Stretta slice decreases towards the south to almost zero, indicating minor displacement along this thrust fault (Fig. 3). Cretaceous deformation structures in the Bernina nappe are mainly related to the D1a event. Deformation involves succession of medium amplitude (m to km) upright to moderately inclined folds leading to a succession of syn- and anti-formal folds with sub-horizontal to gently NE-SW trending axes (Fig. 12). These folds are typically associated with thrust systems (Figs. 7, 9, 12). During the Tertiary (D3 event), the Bernina nappe was overprinted by large amplitude (~km scale) upright folds (F3) associated with sub-horizontal E-W trending axes (Figs. 8, 12).

A major complexity in the Bernina nappe is the occurrence of discontinues slices of remnants of Triassic and Jurassic pre- and syn-rift sediments overlying basement rocks (Fig. 3). Well-exposed examples are the Piz Alv, Piz Mezzaun, Sassalb, Corn Gessi, and Valle Del Monte slices (Figs. 3, 10, 12). These slices were interpreted as preserved east-dipping normal faults associated with tilted blocks (Furrer 1985; Eberli 1988; Froitzheim and Manatschal 1996). This interpretation is supported by the presence of Liassic breccias (Schüpbach 1973; Finger 1978) associated with erosion and tilting of the Lower Jurassic and Triassic pre-rift sequences. However, this interpretation did not explain the local occurrence of pre- and syn-rift

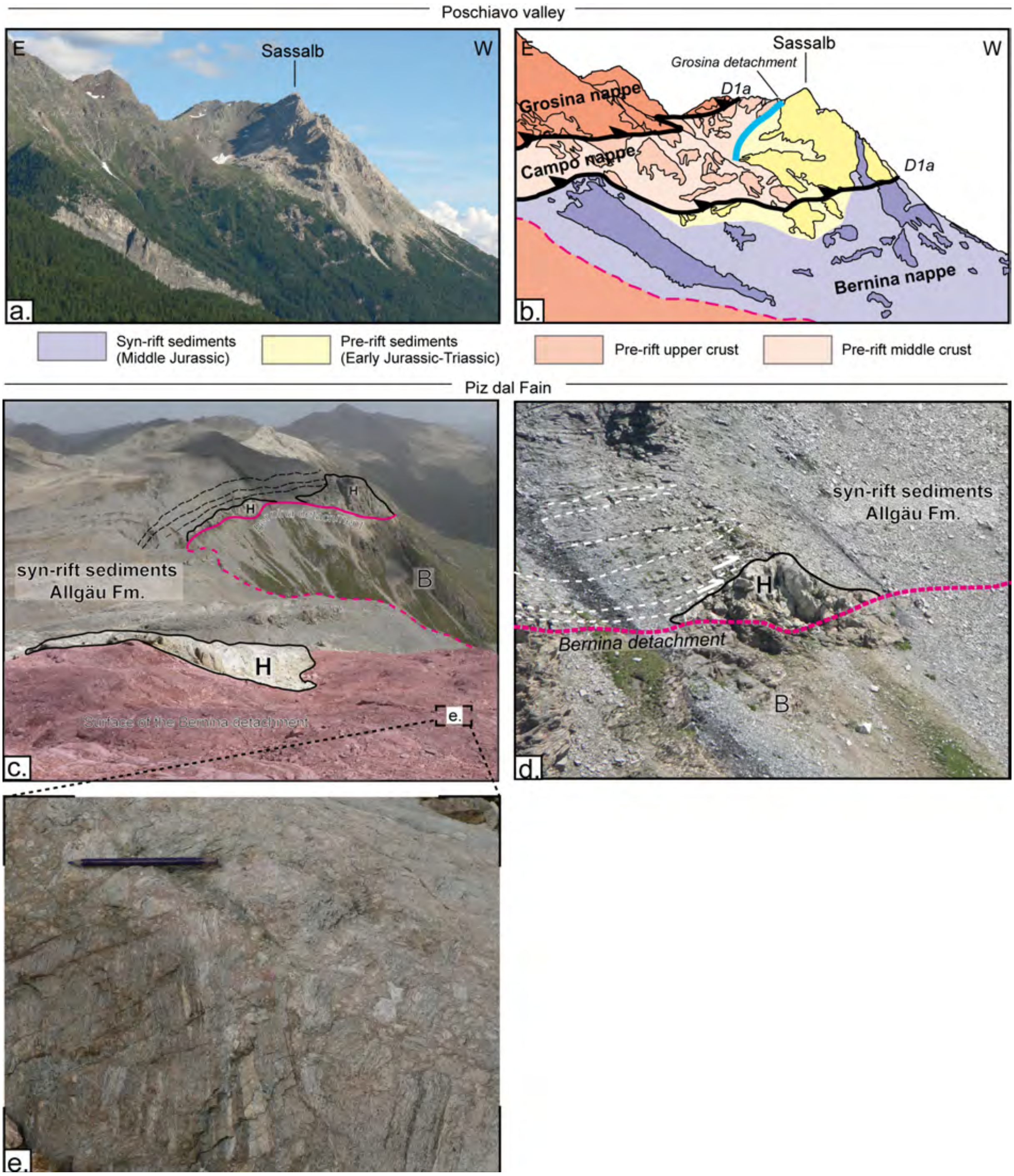


Fig. II. 11: (a, b) Photograph and line drawing from the Sassalb area (for localization of the photograph see Fig. 10) showing the contact between Campo-Grosina and Bernina nappes. Crucial observations are that Bernina are juxtaposed against the Campo nappe before Alpine convergence (blue fault) and these two units were subsequently trusted by the Grosina nappe during the D1a Alpine phase (see text) (804703/135827, Swiss topographic coordinates). (c) View of the Bernina detachment separating a cataclastically deformed basement (B) in the footwall from allochthones of Triassic sediments (H) in the hanging wall. Both basement and allochthones are onlapped by syn-rift of the Allgäu Fm. (794800/149104). (d) Syn-rift sediments onlapped onto allochthones of Triassic sediments (H) or the exhumed basement (B) (795388/149553). (e) Photographs of the Bernina detachment showing cataclastically deformed Paleozoic basement of the Bernina nappe (794805/149104) (for localization of the photograph see Fig. 12).

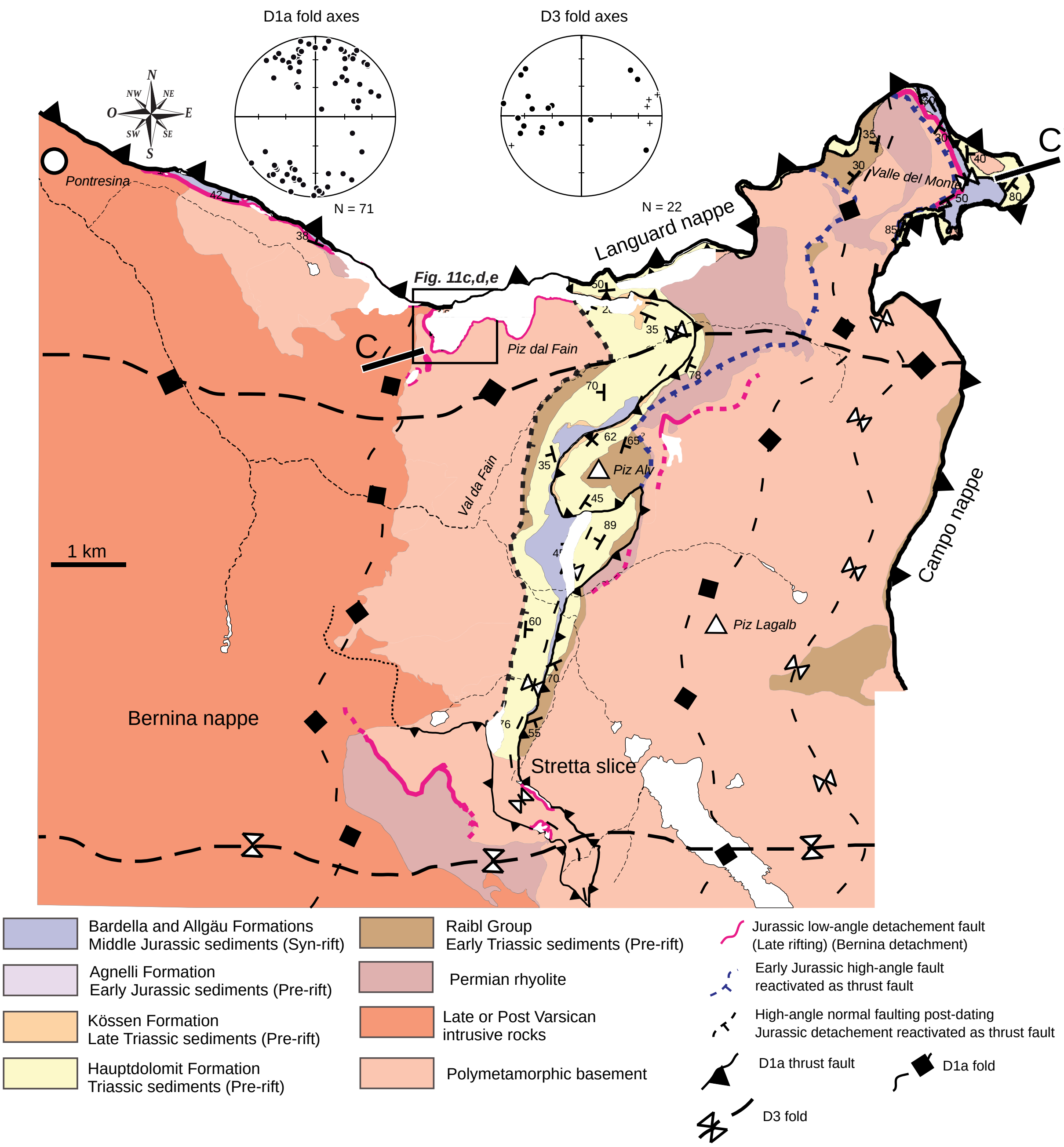


Fig. II. 12: Map of the Val dal Fain – Valle del Monte area showing the distribution of the main Alpine and pre-Alpine structures and their relation to the Mesozoic sediments. Note the reactivation of a Jurassic high-angle normal fault (in blue) during the D1a phase. The stereographic projection is in equal area-lower hemisphere representation.

units. Detailed mapping demonstrates the primary relationship of these slices with syn- and post-rift sequences, as well as their relation to the underlying basement. The complexity of these structures is best preserved along the section across Valle del Monte and Val dal Fain (Figs. 12, 13), where their relationship to Alpine structures is well exposed.

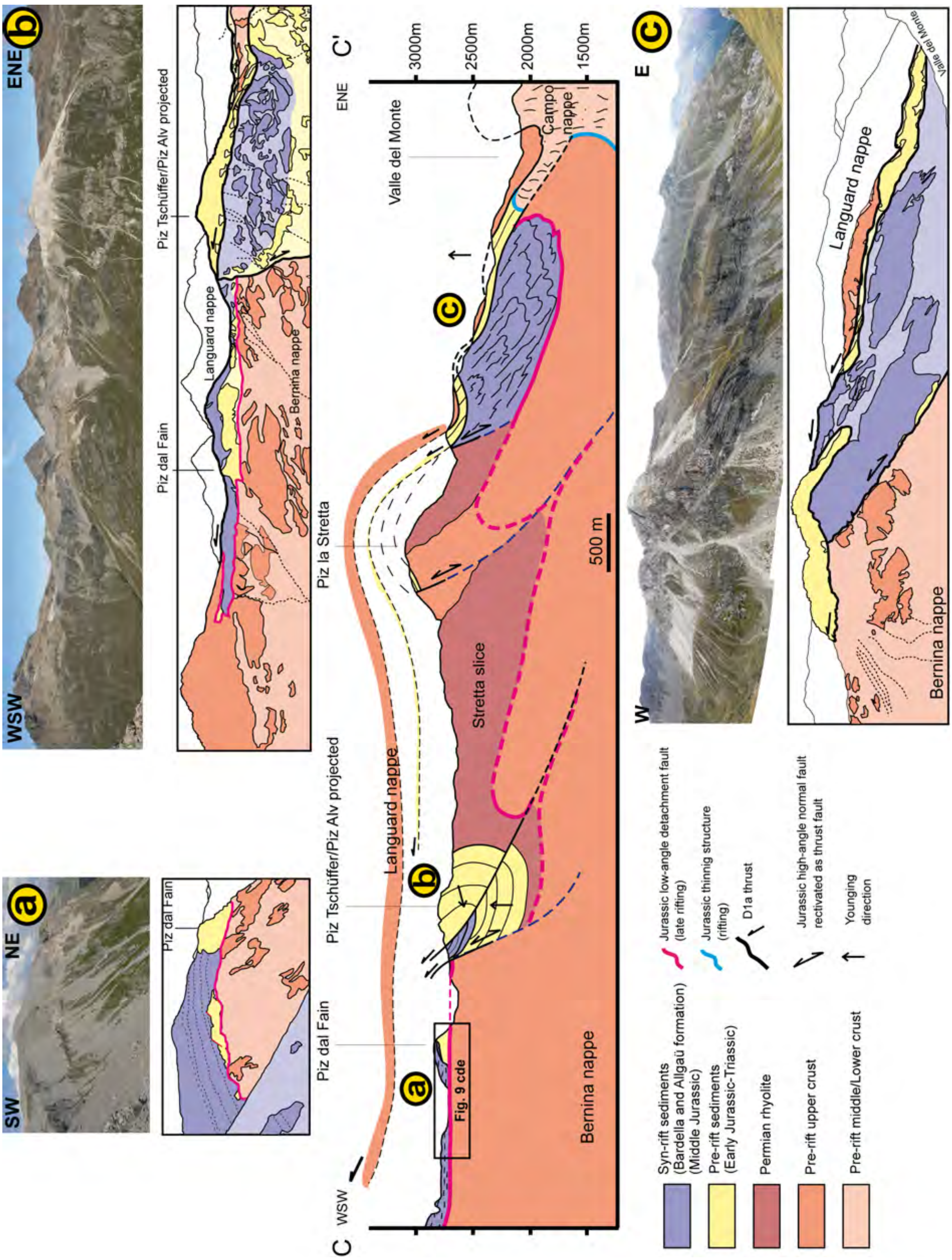
In Valle del Monte in the eastern part of the section, D1a structures are formed by folds and thrusts that overprint the juxtaposition of the Mesozoic cover of the Bernina nappe and pre-rift mid-crustal rocks of the Campo nappe (Fig. 13). The overall structure shows that remnants of the Mesozoic cover of the LAA are preserved in a synform while basement rocks of the MAA (Campo) are exposed in an antiform. A decollement surface along the contact between syn- and pre-rift cover develops westwards into a thrust that ramps up into the syn- and post-rift sediments. To the south at Piz Alv/Piz Tschüffer, the Allgäu Formation overlies the Agnelli limestone dated by ammonites as Sinemurian (Finger 1978). The relationship at Piz Alv/Piz Tschüffer result from recumbent folding associated with a thrust (Figs. 12, 13). Further to the West, the Mesozoic cover is directly juxtaposed with basement rocks along a pre-Alpine Jurassic, high-angle normal fault. In the hanging wall of this fault at Piz dal Fain, syn-rift sediments of the Allgäu Fm. overlie both basement and slices of Triassic dolomites (Fig. 11c,d, 12). In this area, the angle between the onlapping sediments and basement is less than 20°, indicating that the basement had to be exhumed at a low angle (Fig. 11c, d, 12). The contact between Allgäu sediments and the basement has been previously interpreted as an erosional surface of a horst subsequently covered by younger Allgäu sediments (Schüpbach 1973). However, the top of the basement is covered by brittle fault rocks (cataclasites and gouges). Therefore we interpret the top of the basement as an exhumed fault surface that was in turn covered by post-tectonic sediments (Fig. 11c,e). In addition, a high-angle fault displacing the top of the basement is sealed by sediments of the Allgäu Formation. All these observations indicate that the contact between basement and sediments observed in Val dal Fain was not overprinted by Alpine deformation. We interpret this contact as a rift-related detachment fault, similar to the Err detachment (Froitzheim and Eberli 1990; Manatschal and Nievergelt 1997).

Top-basement detachment faults can also be identified and mapped east of Piz Alv (Fig. 12) at the boundary between Permian rhyolites and basement, as well as in the Valle del Monte and at Sassalb. In these two areas, the detachment fault forms the top of the basement and is overlain by sediments of the Allgäu Fm., as in Val dal Fain (Fig. 11c,d, 12). The footwall of this detachment fault comprises a poly-metamorphic basement and Variscan granitoids, characteristic of the pre-rift upper crust. The hanging wall is much more complex, made by remnants of Jurassic tilted blocks bounded by high-angle normal faults (e.g. Piz Alv, Sassalb, Valle del Monte). We propose that the Bernina basement is also covered by top-basement detachment faults that are overlain by extensional allochthons (see discussion below).

6. 3. Alpine and pre-Alpine structures in the Platta and Err nappes

A section across the Err and Platta nappes was published by Manatschal and Nievergelt (1997) (see their Fig. 2). This section shows the same deformation style as the one described in the reminder of the section. Like the Bernina nappe, the Err and Platta nappes can also be subdivided into sub-nappes that are separated by D1a thrust faults and are in turn cross cut and locally reactivated by a D2 normal fault. Such a normal fault can be traced through Val d'Err into Val Julier and further to the Lunghin Pass where it joints the Lunghin- Mortirolo movement zone (see Manatschal and Nievergelt 1997) (Fig. 3). This structure is folded by a D3 fold in the Julier Valley between Bivio and Pass dal Julia. The D3 folds change from south-vergent to north vergent across the Err nappe (Fig. 8). A major structure, which can be mapped across the whole northern study area is the Ela-Albula steep zone (Figs. 3, 7, 8). Froitzheim et al. (1994) and Manatschal and Nievergelt (1997) interpreted this structure as a D1 structure reactivated during the subsequent phases. Although we agree that D1 structures occur in this zone, we reinterpret this structure as a ramp fold overlying a north vergent blind thrust (Fig. 8). Splays of this north vergent thrust can also be observed further to the south, at the base of the Julier subnappe, as indicated by N-vergent thrust directions reactivating top-to-the-west D1a shear zones.

The restoration of the Err and Platta nappes enabled to restore a Jurassic detachment system that is particularly well preserved in the area of Piz Err-Piz Laviner in the Err nappe and in the area of Falotta in the Platta nappe (Froitzheim and Eberli 1990; Manatschal and Nievergelt 1997; Desmurs et al. 2001). The structures described by these authors and their relation to pre-, syn- and post-rift sediments, as well as the fault rocks found along the detachment faults are comparable to the structures observed in the Val dal Fain section (Figs. 12, 13).


SW
NE
a
Piz dal Fain
WSW
ENE
b
Piz dal Fain
Piz Tschüffer/Piz Alv projected
Languard nappe
Bernina nappe
C
C'
Piz Tschüffer/Piz Alv projected
Piz la Stretta
Valle del Monte
Languard nappe
Stretta slice
Bernina nappe
Campo nappe
Fig. 9 cde
500 m
3000m
2500m
2000m
1500m
W
E
c
Languard nappe
Bernina nappe
Valle del Monte
Syn-rift sediments (Bardella and Allgaü formation) (Middle Jurassic)
Pre-rift sediments (Early Jurassic-Triassic)
Permian rhyolite
Pre-rift upper crust
Pre-rift middle/Lower crust
Jurassic low-angle detachment fault (late rifting)
Jurassic thinnig structure (rifting)
D1a thrust
Jurassic high-angle normal fault rectivated as thrust fault
Younging direction

7. Discussion

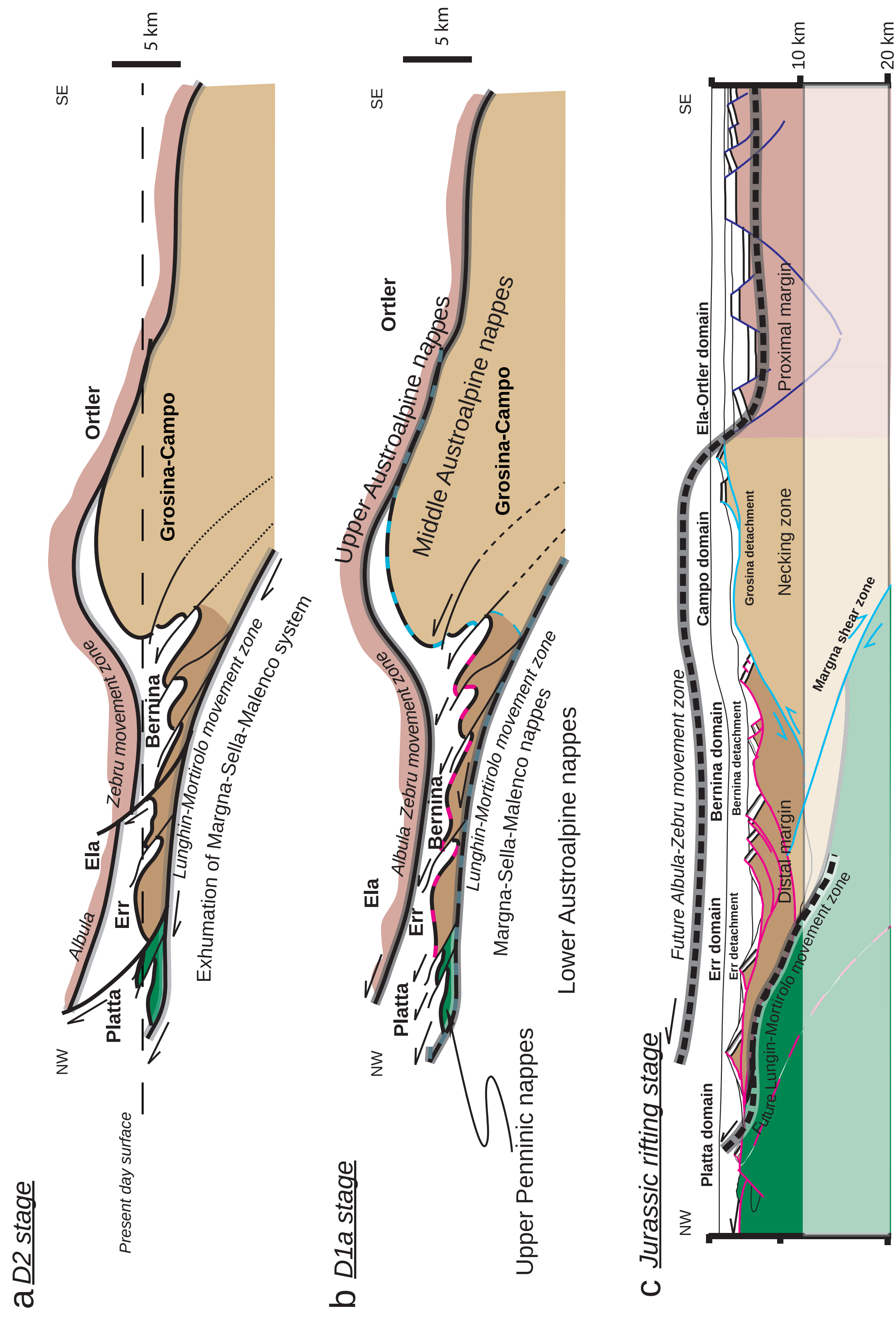
The documentation of rift-related structures in the Bernina, Grosina and Campo nappes highlights the distinct pre-Alpine geometry between the MAA (Grosina/Languard and Campo nappes; necking zone) from the adjacent UAA (Ortler and Ela nappes; proximal margin) and the underlying LAA (Err and Bernina nappes; distal margin). In the following section we discuss the implications of these new observations in the context of reactivation of magma-poor rifted margins.

7. 1. Tectono-metamorphic evolution of the Lower and Middle Austroalpine nappes in SE-Switzerland and N-Italy

The available geological data indicate that the study area has the following characteristics: 1) It has experienced only small displacements along D3 Tertiary N-S directed thrusts; 2) It coincides with a change in the vergence of F3 folds from north- to south-vergent; and 3) the strike of the Adriatic margin was perpendicular to the direction of Tertiary shortening. Thus, it is reasonable to neglect Tertiary N-S deformation in order to restore the geometry of the former margin.

Late Cretaceous extension (D2) was accommodated by various faults and shear zones (e.g. base of Julier normal fault, Handy et al. 1993, Corvatsch shear zone, Spillmann 1993; Mortirolo shear zone, Meier 2003) that flatten out along the Lunghin-Mortirolo movement zone (Figs. 7b, 14). D1a deformation affected the entire area and accounts for the complex fold and thrust belt geometry in the LAA and the UAA. During the D1a phase a W to NW directed nappe stack formed in an external part of the orogen related to the closure of northern part of the Meliata-Vardar domain (Fig. 1). The structures show a succession of syn- and antiformal thrust stacks. From a map view (Fig. 3), the LAA and MAA are made essentially of basement rocks with only local occurrence of Mesozoic cover sequences. In contrast, the UAA (Ela-Ortler nappes) are mainly made of cover sequences. The distribution of cover vs. basement units in the thrust belt seems at odd with the top to the west sense of shear. While in classical fold and

Fig. II. 13: Cross section across the Val dal Fain – Valle del Monte area. Photographs and related line drawings show key observations: (a) Syn-rift sediments (Upper Allgäu Fm. onlaps either onto exhumed basement or pre-rift sediments. (b) Piz dal Fain (view from the S). Jurassic detachment is overlain by blocks of Triassic dolomites (yellow) interpreted as small allochthons. Jurassic syn-rift sediments (violet) seal the contacts. To the east, the detachment is truncated by a former high-angle fault reactivated as thrust fault. At Piz Tschüffer, pre- and syn-rift sediments are folded by D1a deformation. (c) Photograph of Valle del Monte showing the delamination of pre-rift blocks by the Alpine thrust fault and the reactivation of former high-normal fault . Note that the photograph has been inverted to be compared with the map and cross-section shown in Figs. 10, 12.


a D2 stage
Present day surface
NW
SE
5 km
Albula
Platta
Ela
Err
Zebru movement zone
Bernina
Ortler
Grosina-Campo
Lunghin-Mortirolo movement zone
Exhumation of Margna-Sella-Malenco system
b D1a stage
NW
SE
5 km
Ela
Platta
Err
Albula Zebru movement zone
Bernina
Upper Austroalpine nappes
Ortler
Middle Austroalpine nappes
Grosina-Campo
Lunghin-Mortirolo movement zone
Margna-Sella-Malenco nappes
Lower Austroalpine nappes
Upper Penninic nappes
c Jurassic rifting stage
NW
SE
10 km
20 km
Future Albula-Zebru movement zone
Platta domain
Err domain
Err detachment
Bernina domain
Bernina detachment
Campo domain
Grosina detachment
Ela-Ortler domain
Proximal margin
Necking zone
Distal margin
Margna shear zone
Future Lungin-Mortirolo movement zone

thrust belts the proportion of basement to cover sequences increase in the direction of transport, in the study area the opposite is observed. As discussed below, we interpret this as the result of a strong pre-Alpine inheritance. As shown in Fig. 14, the MAA is exposed in an antiformal thrust stack made up of basement, similar to the geometry of the external massifs in the Alpine foreland. We interpret this anticlinal structure to have formed above a major ramp of a west-vergent blind thrust (i.e. Albula-Zebru movement zone). This structure can explain uplift and partial erosion of the MAA as well as the synformal thrust stack and conservation of the sedimentary cover units in the LAA (Fig. 14). This deformation was followed by a more pervasive deformation (D1b), which is well documented in the Margna-Sella-Malenco nappes and may be related to the onset of subduction during the Late Cretaceous.

7. 2. Importance of inherited rift structures during the reactivation of the Alpine Tethys margins

Many papers emphasized the importance of inherited structures in controlling the subsequent compessional overprint. However, most studies either focused to external parts of the orogen and/or the reactivation of classical half graben like rift structures (Handy et al. 1993; Conti et al. 1994; Handy 1996 and Froitzheim and Manatschal 1996). In this paper we discuss the reactivation of a hyper-extended rifted margin. Indeed, the study area is one of the very few areas in the world, where a former distal margin can be restored based on field observations and mapped structures.

In order to understand the importance of rift inheritance and to be able to consider its importance for the subsequent Alpine tectonic overprint, the thermal state, nature, and structure of the crust and lithospheric mantle need to be known for the stage before onset of convergence. Because reactivation of the margin occurred about 70 myr after breakup, we consider that the thermal structure of the rifted margin was equilibrated when Alpine convergence initiated. When considering the nature of the crust, it is important to note that all nappes in the study area preserve basement contacts with pre-, syn- or post-rift sediments, indicating that all these nappes have been near or at the seafloor before onset of Alpine convergence. It is also well documented that the same nappes had to be in the middle crust (Campo nappe), the lower crust

Fig. II. 14: Tectonic, large-scale restoration of the Austroalpine and Upper Penninic nappes in SE-Switzerland and N-Italy. (a) D2 Alpine phase, highlighting the importance of the Albula – Zebru and Lunghin – Mortirolo movement zones in accommodating most of the Alpine deformation. Note also that the relation between the Upper Penninic, Lower and Middle Austroalpine nappes inbetween the two movement zones is only moderately overprinted by Alpine deformation (for restoration see Fig. 14). (b) reactivation of the former rift structures (on the scale of the margin) during the D1 Alpine phase. The Upper Austroalpine, Ela and Ortler nappes are thrust over Middle and Lower Austroalpine nappes. (c) Architecture of the Austroalpine (Adriatic) margin (modified from Mohn et al. 2010).

(Margna nappe) and the subcontinental mantle (Platta and Malenco nappes) during Permian time (see P-T-t paths in Fig. 5 and related discussions). The exhumation cannot be explained by Alpine deformation and had to have occured after Permian but before onset of convergence. Based on primary contacts of the middle and lower crustal and mantle rocks with late Middle and Upper Jurassic sediments (Fig. 3) as well as the available P-T-t paths for these nappes (Fig. 5), it is evident that all these nappes had to have undergone exhumation during Jurassic rifting. This highlights the importance of the rift event in pre-structuring the former rifted margin. Many contacts observed in the LAA and MAA that were interpreted as Alpine, are indeed pre-Alpine or represent reactivated pre-Alpine structures. From the structural point of view, there is strong evidence from the LAA and MAA that some rift-structures were preferentially reactivated and controlled the strain distribution during convergence. This seems to be valid on all scales, from the outcrop to the crustal scale.

On the scale of the whole margin, we suggest that major Alpine structures reactivate pre-Alpine rift-related structures or localize along major lithological boundaries, such as serpentinized mantle, hydrated crust or weak layers in pre-, syn- or post-rift sediments. As an example, during the Alpine subduction (D1b and D2) and continental collision (D3) the Albula-Zebru and in particular the Lunghin-Mortirolio thrust systems were reactivated. We interpret the Albula-Zebru and Lunghin-Mortirolo movement zones as "weak zones" that accommodated most of the strain during compressive and subsequent exhumation stages of the orogen. These zones thus represent major Alpine structures. In the study area these zones can be mapped as complex, highly deformed poly-phase deformation zones, separating 3 main thrust stacks that represent 3 distinguishable paleo-geographic domains of the former margin (Fig. 7).

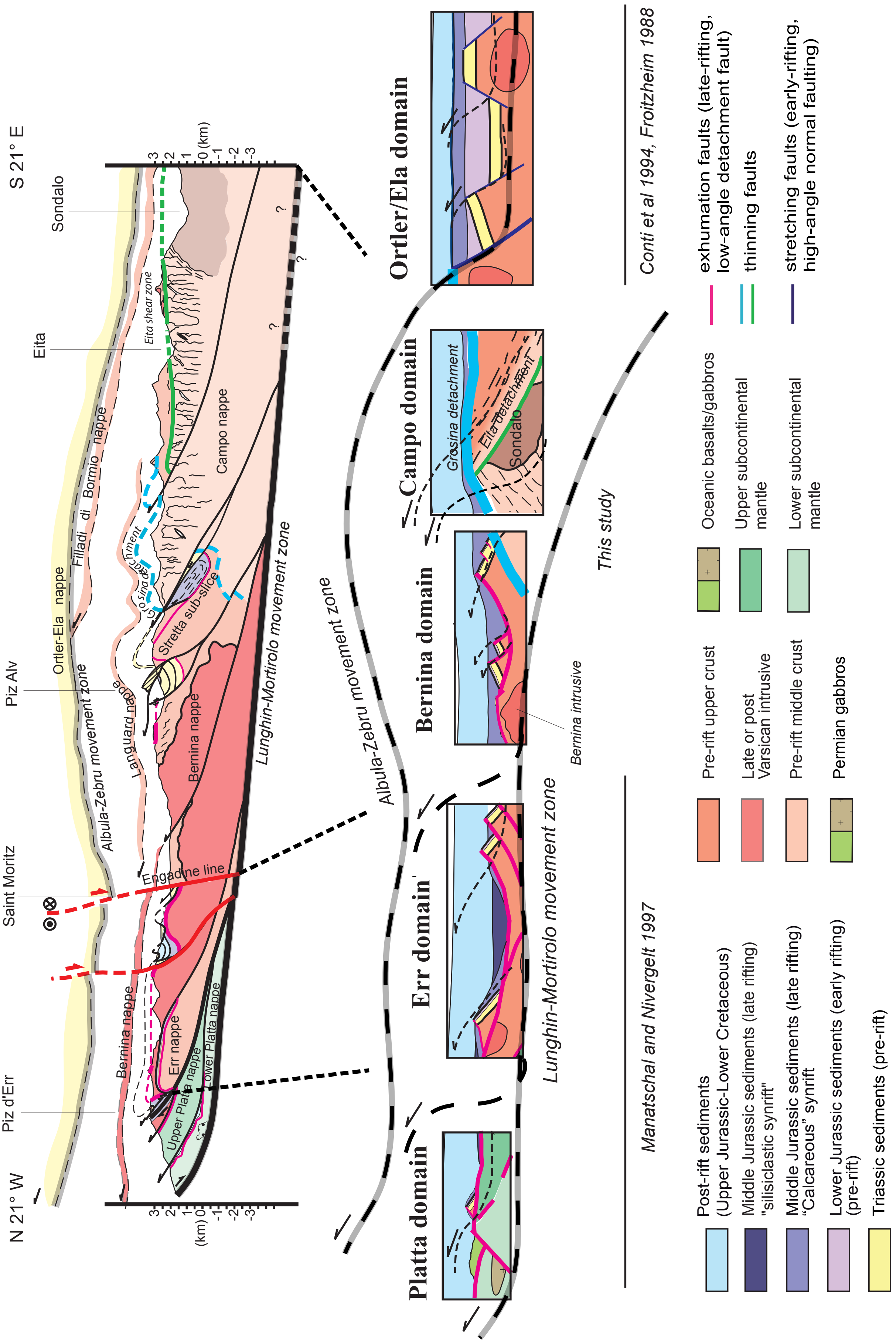
In its present position, the Albula-Zebru movement zone represents the contact between a thrust stack formed mainly by Mesozoic sediments derived from the proximal Adriatic margin that is thrust along a basal decollement over units of the former necking zone and distal margin. The Albula - Zebru fault system ramped from east to west in transport direction from the basement - sediments interface into the Carnian evaporate horizon into the the post-rift sediments. Second order thrust faults were often controlled by inherited east-dipping rift related high-angle normal-faults (Conti et al, 1994). In contrast, west-dipping high-angle normal faults were folded and transported within the nappe stack (Froitzheim, 1988, Froitzheim et al, 1996).

The Lunghin-Mortirolo movement zone represents a poly-phase, complex deformation zone. It separates an upper nappe edifice slightly reactivated from a lower one which shows a stronger Late Cretaceous overprint potentially related to the onset of subduction in the Alpine Tethys margin, as indicated by the pressure dominated metamorphism (Guntli and Liniger 1989). However, the overall paleogeographic context of this area is very complex (Froitzheim et al. 1996; Müntener and Hermann 2001). Handy (1996) and Froitzheim et al. (1996) suggest that the Margna-Sella nappe may represent a former extensional allochthon separated along a mantle window (Platta nappe) (Fig. 3) from the continental margin. Thus, the former geometry of the margin could have included V-shape mantle domains map view, as inferred for the ocean continent transition of the Iberia margin (Péron-Pinvidic et al. 2007). The Lunghin-Mortirolo movement zone may represent the superficial expression of subduction initiation within a mantle window at the contact between exhumed and serpentinized mantle (Malenco) and overlying hyper-extended crust (Margna). This situation is similar to the geometry that has been proposed for the Sesia zone by Babist et al. (2006) and for the eclogitic Piemonte units in the Western Alps (Beltrando et al. 2010).

7. 3. Restoration of the Middle Austroalpine nappes: the missing link between the proximal and distal margin

Methodology and problems of the restoration

Previous studies proposed a kinematic restoration of the UAA (Ela and Ortler nappes) (Conti et al. 1994) and LAA (Err and Platta nappe) (Froitzheim and Eberli 1990; Handy et al. 1993, 1996; Handy 1996; Manatschal and Nievergelt 1997, Desmurs et al. 2001). These studies provided 2- and 3-dimensional restorations of the proximal and the most distal continental margins and its transition into embryonic oceanic crust. In this study we add the missing link between the former proximal and distal margins, which we term the necking zone, exposed in the MAA nappes. In order to simplify the new restoration, we only consider the nappes that are in-between the Albula-Zebru and Lunghin-Mortirolo movement zones (Fig. 9). As discussed above, the main structures within this part of the orogen are D1a structures. A kinematic inversion of these structures reveals the relative positions of the different nappes within an East-West section of the former Adriatic rifted margin. For the reconstruction, two assumptions were made: 1) since the nappes are assembled in a fold and thrust belt, we can assume that the highest position within the nappe stack corresponds to the most proximal position in the former passive margin, and 2) since most units preserve primary contacts to sediments and the individual thrust sheets are less than 3 km thick, all nappes in the thrust stack represent the uppermost few kilometers of the former rifted margin before its reactivation during Alpine shortening. A major limit of the restoration is the lack of absolute E-W shortening. Thus, in our restoration we provide a minimum width of the restored zone. However, as discussed above, we assume


N 21° W
S 21° E
Piz d'Err
Saint Moritz
Piz Alv
Eita
Sondalo
Ortler-Ela nappe
Filladi di Bormio nappe
Albula-Zebru movement zone
Languard nappe
Grosina detachment
Eita shear zone
Stretta sub-slice
Campo nappe
Bernina nappe
Err nappe
Upper Platta nappe
Lower Platta nappe
Engadine line
Lunghin-Mortirolo movement zone
(km)
Platta domain
Err domain
Bernina domain
Campo domain
Ortler/Ela domain
Grosina detachment
Eita detachment
Sondalo
Bernina intrusive
Manatschal and Nivergelt 1997
This study
Conti et al 1994, Froitzheim 1988
Post-rift sediments
(Upper Jurassic-Lower Cretaceous)
Middle Jurassic sediments (late rifting)
"silisiclastic synrift"
Middle Jurassic sediments (late rifting)
"Calcareous" synrift
Lower Jurassic sediments (early rifting)
(pre-rift)
Triassic sediments (pre-rift)
Pre-rift upper crust
Late or post Varsican intrusive
Pre-rift middle crust
Permian gabbros
Oceanic basalts/gabbros
Upper subcontinental mantle
Lower subcontinental mantle
exhumation faults (late-rifting,
low-angle detachment fault)
thinning faults
stretching faults (early-rifting,
high-angle normal faulting)

that most of the shortening was accommodated along the two major movement zones and that the shortening within the Middle and Lower Austroalpine units was minor (e.g < 10km, Fig. 14). Another limitation is that most pre-Alpine contacts were reactivated. However, locally a few basement outcrops preserve primary contacts and give important insights on the pre-Alpine deformation history.

Ela-Ortler domain (proximal margin)(Figs. 14, 15)

This domain is located at the northern and northeastern border of our study area and includes the Ortler and Ela nappes. Froitzheim (1988); Eberli (1988); Conti et al. (1994) described the occurrence of Liassic syn-sedimentary high-angle normal faults associated with tilted blocks affecting a very thick and continuous Triassic carbonate platform. The structures that were restored in this part of the margin correspond to the classical rift structures described in present-day rifted margin such as the North Sea (e.g. Cowie et al. 2005).

Grosina/Languard and Campo nappes (necking zone) (Figs. 14, 15, 16)

The Grosina/Languard nappes consist essentially of basement rocks and preserve only locally primary contacts with pre-rift sediments indicating that these nappes may represent pre-rift upper crust. The presence of the Grosina detachment associated with gouges and cataclasites in an intra-basement position at the top of the Grosina nappe is similar to the Jurassic Err detachment (Manatschal and Nievergelt 1997), suggesting that Jurassic detachment structures can be found on a regional scale (Figs. 3, 10, 11, 12 and 13).

The Campo nappe contains Permian gabbros that were emplaced at a mid-crustal level (Tribuzio et al. 1999; Braga et al. 2001, 2003). $^{40}Ar/^{39}Ar$ data on biotite (Meier 2003) as well as K/Ar ages on biotite and muscovite (Meier 2003; Thöni 1981) on the country rocks of this mid-crustal intrusion record cooling ages around 180-200 Ma. These data indicate exhumation and cooling of these rocks from mid crustal levels during Jurassic rifting (see P-T-t path for Sondalo gabbro in Fig. 5). Staub (1946) described the occurrence of a thin slice of Triassic dolomite along the contact between the Campo and Grosina nappes (Figs 3, 10). While the contact to the overlying Grosina unit is clearly a D1a thrust contact, the contact between the dolomite and the Campo nappe is more difficult to interpret. The contact is characterized by a mylonitic shear

Fig. II. 15: (a) Restoration of the Alpine section across the Upper Penninic, and Austroalpine nappe systems in SE Switzerland and N-Italy. (b) section highlighting the domains from which we can restore the primary architecture of the margin. For the Err-Platta restoration see Manatschal and Nievergelt (1997) and for the restoration of the Ortler/Ela nappes see Conti et al. 1994 and Froitzheim (1988). The restoration of the Bernina and Campo domains are new and discussed in the text.

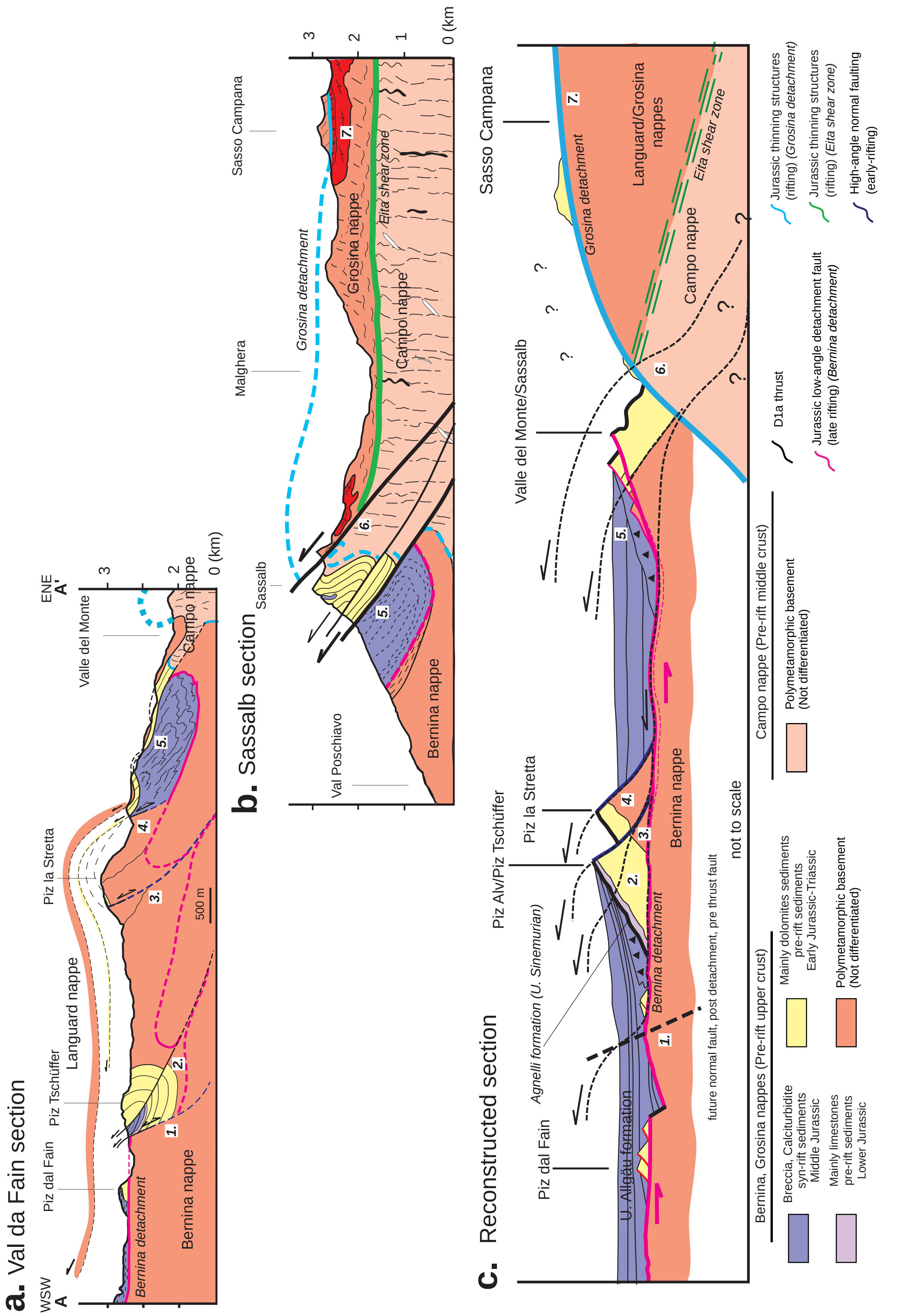


Fig. II. 16: Sections across (a)Val dal Fain and (b) Sassalbo – Val Grosina and (c) their restoration showing the importance of the rift inheritance during the subsequent Alpine tectonic overprint. Note that in many places the Alpine thrust faults reactivate former Jurassic detachment faults.

zone. Although we cannot exclude that this mylonite is of Alpine origin, it is difficult to explain how Triassic dolomites can be thrust onto mid-crustal rocks by Alpine deformation alone. We thus propose that the top of the Campo basement is a Jurassic exhumation structure (i.e. Grosina detachment), which has been reactivated during Alpine convergence. The pre-Alpine nature of this contact is best preserved at Sassalb, where primary relationships are not reactivated by D1a structures (Figs. 10 and 11a,b). In this area we observe the Mesozoic cover (Triassic and early to middle Jurassic sediments) of the Sassalb slice, juxtaposed against the exhumed middle crust of the Campo basement along the Grosina detachment that is made by gouges and cataclasites (Fig. 11a,b). In our restoration (Fig. 16), we propose that the Grosina detachment cuts into mid crustal levels (Campo nappe). During the exhumation in the footwall of a detachment system, the mid-crustal rocks of the Campo basement were juxtaposed against the pre-rift dolomites and syn-rift sediments observed at Sassalb. This interpretation is compatible with Jurassic cooling ages found in the Campo nappe.

Because the Grosina/Languard and Campo nappes are in an Alpine tectonic position between nappes derived from the former proximal (UAA) and distal margin (LAA), and the Ocean Continent Transition (Platta nappe) the structures discussed above developed at the transition between the proximal and distal margin. In present-day margins, this is the area where the crust thins from ±30 km (proximal margin) to < 10 km (distal margin). The domain where the crust is thin in present-day rifted margins is referred to as the “necking zone” (e.g. Lau et al. 2006, Osmundsen and Ebbing 2008). Additional evidence for major crustal thinning is provided by exhumed lower crustal and upper mantle rocks in the Margna-Malenco domain (Fig. 5) (Müntener et al. 2000; Trommsdorff et al. 1993). The distribution of crustal rocks, from mainly upper and mid crustal rocks east of the Poschiavo and Bernina valleys to lower and upper crustal rocks to the west (Fig. 3) is thus inherited from the Jurassic rifting phase. We conclude that the Middle Austroalpine Campo and Grosina/Languard nappes are the remnants of a former necking zone, which accommodated important crustal thinning during Jurassic rifting. In this interpretation, the Campo and Grosina/Languard nappes do not represent the basement nappes of the Mesozoic UAA nappes. This is also supported by the fact that the Campo nappe represents pre-rift middle crust, which excluded primary stratigraphic contacts to Triassic pre-rift sediments (Fig. 10). In this interpretation the Campo nappe can neither be part of the UAA nor the LAA. We thus suggest that the Campo and Grosina/Languard nappes form the MAA that are derived from the former necking zone of the Adriatic rifted margin and represent therefore a distinct paleogeographic domain.

Restoration of the Bernina domain (the distal margin and OCT (Fig. 14, 15, 16))

One major new result reported in this work is the discovery of rift-related detachment faults similar to those reported from the Err detachment in the Err nappe further to the north (Manatschal and Nievergelt 1997). The footwall of the low-angle detachment system in Val dal Fain (Bernina nappe) is formed by Late Variscan to Permian intrusive rocks emplaced at shallow crustal levels. The observation that the detachment fault is associated with pre-rift upper crustal rocks indicates that this structure was active at shallow crustal levels. The hanging wall of the detachment system is characterized by extensional allochthons consisting of pre-rift upper crust and a pre- to syn-rift sedimentary cover. These allochthons formed by delamination of the hanging wall during exhumation of continental crust along downward concave detachment faults. This process leads to the creation of exhumed fault surfaces that are onlapped by syn-rift sediments and sealed by pelagic post-rift sediments (Figs. 11c,d,e, 13). Such contacts are documented in the Val dal Fain (Figs. 11c,d,e, 12, 13). Top basement detachment faults associated with extensional allochthons highlight the major difference between the rift structures forming the distal margin and those forming the proximal margins. The change in deformation style has also important implications for the reconstruction of the LAA and MAA.

The Mesozoic cover in the UAA shows a continuous Triassic carbonate platform affected by Liassic syn-sedimentary high-angle normal faults filled with early Jurassic syn-rift sediments (Froitzheim 1988; Eberli 1988; Conti et al. 1994). The pre-Alpine geometry is similar to classical rift structures described in present-day rifted margins such as the North Sea (e.g. Cowie et al. 2005), whereas the LAA displays a geometry made by discontinuous slices of Mesozoic cover surrounded by basement rocks (Figs. 3, 13). In the past, the Jurassic geometry was interpreted as a succession of tilted blocks similar to the rift geometry reported from the UAA (e.g. Finger 1978, Eberli 1988; Handy et al. 1993). However, since we find onlapping syn- and post-rift sequences on the basement as well as low-angle detachment faults capping the basement (e.g. Val dal Fain; Figs. 11c,d,e, 12, 13), we propose that exhumed basement and extensional allochthons are the rule rather than the exception in the LAA. This is also compatible with observations made in present-day distal rifted margins, where the lack of high-angle faults and tilted blocks is observed and supported by deep sea drilling (e.g. Manatschal et al. 2001; Reston et al. 2009). We therefore suggest that the geometry of Mesozoic slices of sediments surrounded by basement is inherited from Jurassic rifting, where the pre-rift cover has been dismembered on the hanging wall of a low-angle detachment fault leading to the exhumation of basement and the creation of a new surface sealed by syn- and post-rift sediments.

The importance of inherited rift structures and, in particular, of the necking zone in localizing deformation during inversion of the margin is also observed in present-day rifted margins. Cunha et al. (2010) demonstrate for the present-day SW Portugal rifted margin that the thrust front resulting from reactivation of the passive margin is located at the transition from the distal to the proximal margin in the necking zone. Therefore we conclude that Austroalpine nappe systems represent an excellent natural laboratory to study the reactivation processes of hyper-extended magma poor-rifted margins.

8. Conclusion

The documentation of rift related structure in the Middle Austroalpine nappes (Campo, Grosina/Languard) and their relations to the Lower and Upper Austroalpine nappes yields new constraints for the paleogeographic and Alpine evolution of the Austroalpine domain. The strong pre-Alpine (mainly Jurassic) inheritance determined the complex geometry of the Austroalpine nappe stack. Based on new mapping and the construction of Alpine cross-sections, we propose the following new interpretations:

- The Campo, and Grosina/Languard nappes represent relics of a necking zone of the Early to Middle Adriatic rifted margin as indicated by the occurrence of intermediate crustal rocks juxtaposed with Mesozoic pre-rift Triassic dolomites along Jurassic detachment faults (e.g. Grosina detachment, Eita shear zone in Figs. 3, 9, 10).
- The Bernina nappe represents a relic of the Adriatic distal margin with low-angle, top-basement detachment faults and extensional allochthons that were partly reactivated during Alpine convergence.
- The major structures accommodating strain during the polyphase evolution of the Austroalpine nappe systems were the Albula-Zebru and Lunghin-Mortirolo movement zones, resulting from reactivation of the necking zone.

In conclusion, the restoration shows that some of the complexities in collisional orogens are derived from the reactivation of inherited rift structures. The close link between Alpine and pre-Alpine structures, particularly in distal margins, is fundamental for understanding the tectonic and rheological evolution during reactivation and for the emplacement of the remnants of the former rifted margin within a mountain belt.

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TROISIÈME PARTIE

Le chapitre second traite des processus menant à l'amincissement crustal lors du rifting. Ce sujet est un complément essentiel au chapitre précédent traitant de la caractérisation de la géométrie Alpine et de la restauration de la situation pré-Alpine dans les reliques de la marge distale (nappe de Bernina) et dans la « necking zone » (nappes de Campo/Grosina). L'objectif ici est de caractériser et de contraindre l'architecture crustale de la « necking zone » et de la marge distale ainsi que celle des structures actives pendant l'amincissement lors du rifting. L'analyse s'appuie sur des données de terrain acquises pendant trois saisons (2007, 2008, 2009) dans les Alpes au Sud-Est de la Suisse et le Nord de l'Italie. Les observations ont été complétées par une étude structurale et micostructurale ainsi que par des datations obtenues par la méthode Ar/Ar. Pour ces dernières, j'ai effectué moi-même les séparations des minéraux à Strasbourg (EOST), tandis que les mesures d'âge ont été faites par Richard Spikings à l'université de Genève.

Ce chapitre est composé de cinq parties :

La première partie décrit l'architecture générale de nombreuses marges passives actuelles. Ces observations sont couplées à une synthèse des modèles proposés de l'amincissement crustal durant le rifting.

La deuxième partie décrit succinctement des reliques de la marge Adriatique dans les Alpes Centrales et du Sud. Cette description est complétée par celle de la réactivation Alpine subie par la marge Adriatique dans les Alpes Centrales, région de mon étude.

La troisième partie traite de la description et de la caractérisation de l'architecture des zones de Campo/Grosina et de Bernina interprétées comme représentant respectivement les reliques de la « necking zone » et de la marge distale Adriatique.

La quatrième partie est dédiée à l'étude des structures tectoniques préservées dans les nappes de Campo/Grosina et de Bernina actives pendant le rifting et ayant pu accommoder l'amincissement crustal.

La cinquième partie est une discussion intégrant toutes les observations présentées dans les parties précédentes. Cette synthèse permet de proposer un modèle conceptuel du processus d'amincissement de la lithosphère continentale pendant le rifting. Les résultats sont comparés aux exemples de marges passives actuelles.

Cette section du mémoire de Thèse fait l'objet d'un article qui sera soumis pour publication à la revue Tectonics.

Publication 2: in preparation for Tectonics

HOW DOES THE CONTINENTAL CRUST THIN DURING RIFTING IN MAGMA-POOR RIFTED MARGINS: EVIDENCE FROM THE BERNINA/CAMPO/GROSINA UNITS IN THE CENTRAL ALPS (SE-SWITZERLAND AND N-ITALY) AND IMPLICATIONS FOR PRESENT-DAY RIFTED MARGINS

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ABSTRACT

From many present-day magma-poor rifted margins it could be observed that the transition from little thinned continental crust in proximal margins to hyper extended crust (≤10 km) in distal margins occurs within a well-defined necking zone. The low resolution of the available offshore data and the lack of drilling make it difficult to study the structures and processes associated with crustal thinning in present-day rifted margins. A more direct access to the stratigraphic record and crustal architecture of deep-water rifted margins is exposed in the Alps in Western Europe. We focused our study to the Austroalpine Bernina-Campo-Grosina units preserving relics of the former necking zone of the Alpine Tethys rifted margin. Within this necking zone, different high-strain shear zones responsible for lithospheric thinning can be described including: 1) a system of conjugate low angle shear zones/detachment faults active in the brittle upper crust (Grosina detachment) and lower crust (Pogallo type shear zone); 2) mid-crustal decollements decoupling the deformation in the upper and lower crust (e.g. Eita shear zone); and 3) an extraction shear zone, whose activity resulted in the total excision of the middle crust (e.g. Margna shear zone). These high-strain zones are interpreted to accommodate the crustal thinning from 30 to 10 km during Piensbachian to Toarcian time (± 180Ma). Thinning resulted in exhumation of mid-crustal rocks in the necking zone, while in the distal margin upper and lower crust are juxtaposed and overprinted by late detachment faults that cut across the thinned crust and exhume mantle rocks to the seafloor. These structures can explain the first order crustal architecture observed at many present-day rifted margins.

1. Introduction

The discovery of exhumed subcontinental mantle at Ocean-Continent Transitions (OCT) at present-day rifted margins leads to the question of how does the continental crust thins at hyper-extended, magma-poor rifted margins. Understanding the processes related to extreme crustal thinning is fundamental to unravel the structure and evolution of rifted margins. In OCT and deep-water rifted margins, the final extension is demonstrated to be accommodated by long offset, downward concave detachment faults leading to exhumation of lower crustal and subcontinental mantle rocks at the seafloor (Whitemarsh et al. 2001). As shown by numerous previous authors (Manatschal 2001; Pérez Gussinyé and Reston 2001; Reston 2007), when the downward detachment faults became active, continental crust was already thinned to less than 10 km. However, the amount of visible extension along normal faults is insufficient to explain this extreme crustal thinning (Kusznir and Karner 2007; Reston 2007). Therefore major crustal thinning must have occurred previously and by a different process. Observations from reflection and refraction seismic data are showing that the major crustal thinning is located in the so-called necking zone, which marks the transition from ”original” crustal thickness (± 30 km) at the proximal margin to hyper extended crust (≤ 10km) at the more distal part (Péron-Pinvidic and Manatschal 2009). In recent years most studies on rifted margins focused on mantle exhumation or excess magma production, both of which occur in final stages of rifting and result in either magma-poor or magma-rich rifted margins. However less attention has been paid on the processes that can explain crustal thinning predating mantle exhumation or excess magma extrusion. Such processes are at present poorly constrained, mainky due to the limitations of seismic and other geophysical methods in revealing the nature of the rocks and deformation processes in the necking zones as well as the lack of drill hole data penetrating basement in present-day rifted margins, make that processes responsible for crustal thinning are at present poorly constrained. Therefore we focus our study on the ancient Alpine Tethys margins in the Alps. Remnants of the ancient Alpine Tethys Jurassic rifted margin are well preserved and exposed and the palaeo-geographic position of these units can be restored with relative confidence (Froitzheim et al. 1994; Froitzheim and Manatschal 1996; Mohn et al. 2010). The aim of this study is to unravel and understand the major crustal thinning documented in the Bernina-Campo-Grosina units, which sampled the transition between the former distal and proximal margins of the Adriatic rifted margin. The main questions addressed in this study are: where and when did crustal thinning occur, what are the structures and processes that can explain such a major crustal thinning and under what conditions were these structures actives? In this paper we will present new structural, petrological and thermochronological data in order to constrain the crustal structure of this fossil necking zone. These data will permit to discuss the processes controlling crustal thinning and to propose a coherent model to explain the structural and temporal evolution of magma-poor rifted margins.

2. Structure and evolution of magma-poor rifted margins

The classification of rifted margins is mainly based on the volume of magmatism during rifting. Two end members can be described: magma-rich and magma-poor rifted margins. These two types can be differentiated by (1) the volume of syn-rift magmatism, (2) top basement morphology, (3) subsidence history, (4) crustal thickness, and (5) sedimentary/magmatic architecture. However, this subdivision into two "types" of margins underscores the fact that each margin results from the interplay of inheritance (structural, thermal or compositional) and rift-history (sediment supply, vicinity of hot spot, strain rates). However despite of this specificity, it is still possible to recognize characteristic features and structures (see buildings stones in Péron-Pinvidic and Manatschal in press) in rifted margins. In this paper, we will focus on magma-poor rift systems, although we think that extreme crustal thinning, as discussed in this paper, may also occur in magma-rich systems (e.g. Møre margin; Osmundsen and Ebbing, 2008).

2. 1. Architecture of magma-poor rifted margins

Observations from many rifted margins worldwide show that in sections going from the continent towards the ocean a change in the style of deformation and architecture of rift basins can be observed. Within such sections, different domains can be characterized, which are referred to as the proximal margin, the necking zone, the distal margin and the ocean continent transition (OCT) (Fig. 1).

2. 1. 1. Proximal margin

Extension in proximal margins is mainly accommodated by tilted blocks related to fault-bounded rift basins (e.g. Jeanne d'Arc, Tankard et al. 1989, Kusznir et al. 1991; Cowie et al. 2005 and references therein). β-factors are typically less than 2. Extension is accommodated by listric faults, which results in the tilting of the pre-rift sequence, the thickening of the syn-rift sediments into the footwall and the sealing by post-rift sediments. Reflection and refraction seismic data show that the faults sole out at the brittle-ductile transition at around 10-15 km. Thus, deformation in the upper crust and in the upper mantle is decoupled within ductile layers in the middle and/or lower crust depending on initial Moho temperature and composition of the lower crust. This is particularly well shown in the section across Grand Banks (Newfoundland margin) where reflection and refraction seismic investigations show that beneath the Jeanne d'Arc basin, continental crust is still 30 to 37 km thick (Keen and de Voogd 1988; Laun et al. 2006; Van Avendonk et al. 2006). In these areas exhumation of deeper crustal rocks is limited to fault scarps of high-angle normal faults. Proximal margins can be recognized on both conjugate rifted margins, in magma-rich and magma-poor systems, and the age of the rift-structures found in the proximal margin is commonly predating the age of breakup (e.g. Iberia-Newfoundland rifted margin; Manatschal and Bernoulli 1999; Wilson et al. 2001).

2. 1. 2. Necking zone

A characteristic feature in many rifted margins such as the Iberia, mid-Norway and Angola margins is the occurrence of a necking zone here referred to as a zone of extreme crustal thinning that separates the proximal from the distal margin, hereafter referred to as «necking zone». This zone juxtaposes crusts that are little of not thinned (proximal margin) against strongly thinned, less than 10 km thick crust (Fig. 1). This abrupt crustal thinning occurs over a relative narrow area of 50-60 km (Olaffson et al. 1992; Dean et al. 2000; Contrucci et al. 2004; Moulin et al. 2005; Fernandéz et al. 2005 and Osmundsen and Ebbing 2008) and is associated with a change of the Moho geometry from relatively flat in the proximal margin to a dip up to 35° in the necking zone (Lau et al. 2006). Within this domain, β-factors are typically > 2 and (1-1/β) values are < 0.5 (Reston 2009). In many rifted margins, the necking zone is developed in a different way on the two conjugate margins, indicating an asymmetry of the margin (Péron-Pinvidic and Manatschal 2009).

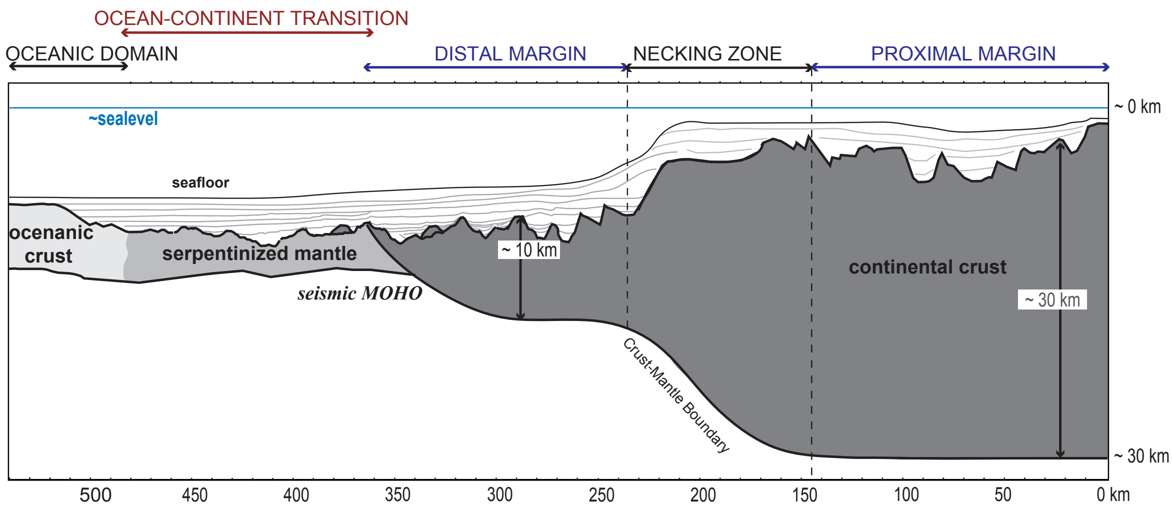


Fig. III. 1: Idealized cross-section across a magma-poor rifted margin showing the different domains, crustal architecture and terminology used in this paper.

2. 1. 3. Distal margin and Ocean-Continent Transition (OCT)

From many rifted margins such as the Iberia-Newfoundland, S-Atlantic and Australian rifted margins (Van Avendonk et al. 2009; Dirscoll and Karner 1998), it can be shown that distal parts are characterized by hyper-extended crust that is separated from the first oceanic crust by a domain of exhumed subcontinental mantle, also referred to as Ocean Continent Transition (OCT). The domain with hyper extended crust is observed in many present-day rifted margins to occur over areas as wide as 200 km (e.g. Unternehr et al. in press). However the amounts of extension that can be inferred from imaged faults are insufficient to explain this extreme crustal thinning, which resulted in a debate about the processes responsible for crustal thinning, also known as the extension discrepancy (Sibuet 1992; Driscoll and Karner 1998, Reston 2007; Kusznir and Karner 2007 and references therein). Hyper-extended continental crust is commonly associated with sag basins (e.g. Angola margin, Moulin et al. 2005; Aslanian et al. 2009), whose the architecture and evolution are yet poorly constrained. Drill holes in seismically imaged hyper-extended crust exist only from the Iberia margin. Based on these data Witmarsh et al. (2001) and Péron-Pinvidic and Manatschal (2009) described the transition from high- to low-angle normal faulting at the edge of the continental crust. However, Manatschal et al. (2001) and Peréz-Gussinyé and Reston (2001) have demonstrated that the structures observed in the deep Iberia rifted margin post-date crustal thinning to less than 10 km. Thus extreme crustal thinning had to occur further inboard, in the necking zone, and it had to predate mantle exhumation at the seafloor.

2. 2. Models to explain extreme crustal thinning

In the late 70's and mid-80's, two end-member models were proposed to describe and explain rift geometries as well as lithospheric thinning and related subsidence history. These two "end-member" models are: (1) the uniform depth-independent pure shear model of McKenzie (1978) and (2) the depth dependent, asymmetric simple shear model of Wernicke (1985) (Figs. 2a and b). However, none of them are able to integrate all observations made at present-day and fossil rifted margins and to explain extreme crustal thinning and the related isostatic evolution of rifted margins. The depth dependent pure shear model, characterized by a uniform stretching with depth (βcrust = βmantle) was initially developed and proposed by Mc Kenzie (1978). This model has been used with success in proximal margins where the crust and lithosphere are not strongly thinned and ß values are typically below 2. However in domains of hyper-extended crust (ß > 2), this model is unable to predict the observed subsidence history as well as the exhumation of mantle rocks and the observed asymmetry of the distal rifted margins. In contrast the simple shear model of Wernicke (1985) proposes a detachment fault cutting across the entire rheologically stratified crust leading finally to mantle exhumation (Fig. 2b). However, this model is however, mechanically unrealistic since detachment faults are unlikely to cut through ductile layers and to accommodate strain along one single structure on the scale of the lithosphere. Numerous studies have emphasized that depth-dependent stretching (DDS) becomes an important process at high ß values (> 2) (e.g. Kusznir and Karner 2007) (Figs. 2c and d). At present, many models have been proposed, which combine pure and simple shear extension and include DDS. These models can be subdivided in DDS-continuous non-uniform stretching models (e.g. Rowley and Sahagian 1986), DDS associated with lower crustal flow (Brun and Beslier 1996; Driscoll and Karner 1998), and DDS coupled with a divergent flow field (Davis and Kusznir 2004; Kusznir et al. 2005) (Fig. 2c). These models can account for most of the observations made along rifted margins; however, none of these models can explain the whole evolution observed along one conjugate rift system. While some models propose flow out and into the rift zone during rifting, suggesting that thinning is controlled by ductile layers on a crustal scale (e.g. Brun and Beslier 1996, and Driscoll and Karner 1998), others suggest that the crust is thinned by polyphase normal faulting (Reston 2007) (Fig. 2d). The key problem that remains in many DDS models is that the rheological evolution of the crust and mantle during thinning is not constrained.

In order to explain the complete evolution of rifted margins, polyphase models had first to be conceptualized ((Lister et al. 1986, 1991; Whitmarsh et al. 2001, Reston et al. 2007). In a later stage they were numerically modeled (Nagel and Buck 2004, Lavier and Manatschal 2006, Huismans and Beaumont 2007) (Fig. 2e). These models can predict the initial and final parts of the rift process observed at magma-poor rifted margins, i.e. they can change from decoupled distributed (pure shear) to coupled and localized deformation (simple shear) and they can explain the major localized thinning of the crust from 30 to 10 km observed at rifted margins. However, the processes that are proposed in these models to localize the deformation and thin the crust remain poorly constrained. This is mainly a consequence of the lack of data. Geophysical and experimental data alone are not sufficient to test these models. Therefore, in order to constrain the crustal evolution during extreme crustal thinning, we will focus our study on the remnants of the ancient Alpine Tethys rifted margins preserved in the Alps. These remnants constitute at the moment the best-documented example of a fossil magma-poor rift system worldwide. This natural laboratory enables us to identify and quantify the crustal architecture in hyper-extended basins as well as to investigate the necking zone based on field mapping, structural, petrological and thermochronological methods.

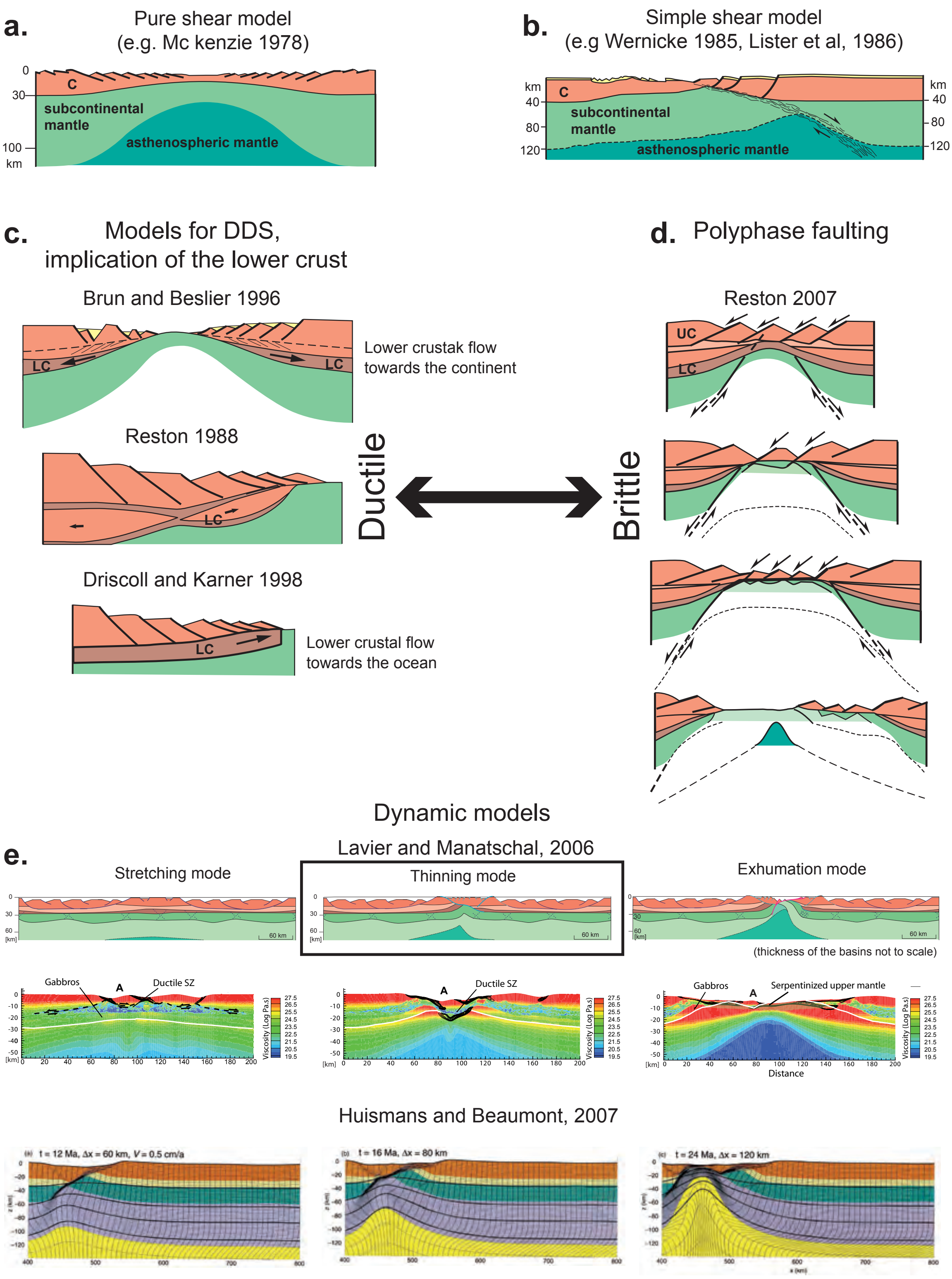


Fig. III. 2: Summary of the principal models and concepts explaining the evolution of rifted margins: (a) Pure shear models; (b) simple shear models; (c) depth dependent stretching (DDS) models; (d) polyphase faulting models; and (e) numerical models (above: conceptual and numerical model evolution as proposed by Lavier and Manatschal (2006) and below the numerical model proposed by Huismans and Beaumont (2007) (for explanation see text).

3. Adriatic margin in the Alpine Tethys domain: a fossil analogue of a magma-poor rifted margin

The evolution of the Alpine Tethys rift system was contemporaneous and kinematically linked with the opening of the Central Atlantic in Jurassic time. It resulted in the separation of Europe and Iberia from Africa and Adria (Figs. 3a and c). From Late Cretaceous onwards the Alpine Tethys began to close resulting in a final collision between Europe and Adriatic in Eocene time. Thus, the Alps in Central Europe resulted from the collision of former rifted margins, which explains that relics of the former European and Adriatic margins are preserved in this mountain belt (e.g. Froitzheim et al. 1994; Beltrando et al. 2010). In this paper we will mainly focus on the fossil Adriatic margin, which is preserved in the Austroalpine and South Penninic nappes in SE-Switzerland and N-Italy (Fig. 3b). Due to their particular paleogeographic position and Alpine evolution (see Mohn et al. 2010) these units underwent a relative mild Alpine tectonic and metamorphic overprint, which makes that the Jurassic rift structures are particularly well preserved. This enabled to propose restored sections across the former Alpine Tethys margins (e.g. Lemoine et al. 1986, Bertotti et al. 1993; Froitzheim and Manatschal 1996; Manatschal et al. 2007). In order to not lose the reader with too many details about the Alpine evolution, we will briefly summarize the present-day architecture and the reconstructed paleogeographic position of the different units of the margin across two sections, one exposed in the Eastern Alps and across the Southern Alps (Fig. 3d) (for a more exhaustive review see Bertotti et al. 1993, Froitzheim et al. 1994, Schmid et al. 1996, 2004, Mohn et al. 2010).

3. 1. The Eastern Alps section

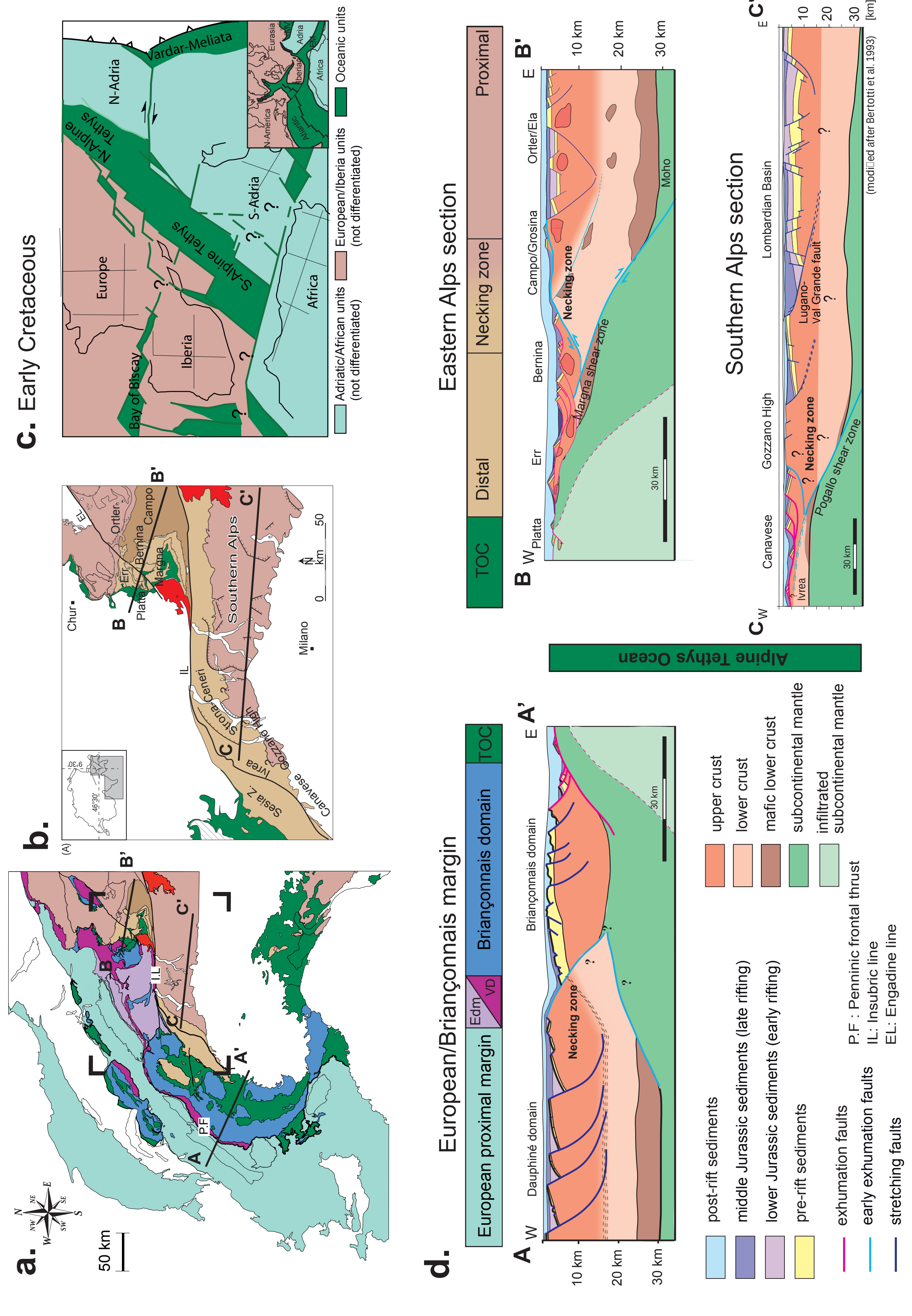
The Austroalpine and South Penninic nappes in SE-Switzerland and N-Italy were emplaced during Late Cretaceous west directed convergence. During this early stage of convergence, more proximal parts of the margin were thrust over more distal parts of the northern Adriatic margin. Thus, the present-day nappe pile can be restored in an east to west section across the rifted margin, which includes remnants from the proximal margin (Upper Austroalpine nappes: Ortler/Ela units), the necking zone (Middle Austroalpine nappes: Campo/Grosina units), distal margin (Lower Austroalpine nappes: Bernina/Err/Margna units) and the OCT (South Penninic nappes: Platta/Malenco units) (Figs. 3b and d and Fig. 4).

In the Ortler and Ela units preserving the proximal margin, the pre-rift sequence overlies Paleozoic basement and consists mainly of continental deposits that grade upsection into shallow marine carbonates (Fig. 3d). Two distinct carbonate platforms of respectively Middle and Upper Triassic age can be distinguished. Relics of Jurassic high-angle normal faults can be found associated with half-graben basins. These fault-bounded rift basins were active during initial rifting from late Triassic to Pliensbachian-Toarcian time and were mainly filled by mass flow breccias and calciturbidites interleaved with hemi-pelagic limestones (e.g Allgäu Formation of Eberli 1988). All reworked sediments were exclusively derived from the pre-rift carbonate platform (Froitzheim 1988).

In the Err and Platta units preserving remnants of the former distal margin and OCT, rift-related low-angle detachment faults are the most characteristic rift structure. These faults separate continental crust and exhumed subcontinental mantle in the footwall from extensional allochthones made of basement and pre-rift sediments. Syn- and post-rift sediments, further oceanwards also extrusive mafic rocks, overlie the detachment faults with an angle of less than 30°, which confirms the relative flat geometry of these faults and their exhumation at the seafloor. In these domains "pre-rift" sedimentation lasted until the Early Jurassic, as indicated by a starved carbonate platform, made of fossiliferous limestone interbedded by shales, marls and dolomites. This platform is capped by a hardground that formed in an open marine environment (Schüpbach 1973). Based on ammonites (e.g. Eparietites *sp.*, Paltechioceras *sp.*), a Sinemurian age can be inferred for the age of the deposition of the carbonates and an early Pliensbachian age (188 Ma) for the formation of the hardground that pre-date onset of extension in the distal margin (Manatschal et al. 2007).

3. 2. The Southern Alps section

A section similar to the one reconstructed and described from the Austroalpine and South Penninic units across SE Switzerland can also be restored across the Southern Alps (Figs. 3b and d). Preserved relics of the former proximal Adriatic margin are exposed in the Lombardian Alps. In this area, Bernoulli (1964), Winterer and Bosellini (1981) and Bertotti et al. (1993) described several high-angle normal faults associated with rift basins, which are contemporaneous from those described in the Eastern Alps (Bertotti et al. 1993 and references therein) or on the European margin (Lemoine et al. 1986). Relics of the former distal margin are exposed in the Canavese and Ivrea-Verbano-Strona-Ceneri zones (Handy et al. 1999 and references therein; Ferrando et al. 2004). Although less well exposed, the structures and evolution of these units are comparable with those observed in the Austroalpine nappes.


a.
50 km
b.
c. Early Cretaceous
Europe
Iberia
Bay of Biscay
Africa
N-Adria
S-Adria
N-Alpine Tethys
S-Alpine Tethys
Vardar-Meliata
Adriatic/African units (not differentiated)
European/Iberia units (not differentiated)
Oceanic units
Southern Alps
Milano
Chur
d.
European/Briançonnais margin
European proximal margin
Edm
VD
Briançonnais domain
TOC
Dauphiné domain
Necking zone
10 km
20 km
30 km
Eastern Alps section
Distal
Necking zone
Proximal
Platta
Err
Bernina
Campo/Grosina
Ortler/Ela
Margna shear zone
Moho
Alpine Tethys Ocean
Southern Alps section
Canavese
Ivrea
Gozzano High
Lombardian Basin
Lugano-Val Grande fault
Pogallo shear zone
30 km
post-rift sediments
middle Jurassic sediments (late rifting)
lower Jurassic sediments (early rifting)
pre-rift sediments
exhumation faults
early exhumation faults
stretching faults
upper crust
lower crust
mafic lower crust
subcontinental mantle
infiltrated subcontinental mantle
P.F : Penninic frontal thrust
IL: Insubric line
EL: Engadine line
(modified after Bertotti et al. 1993)

The Ivrea-Verbano zone is interpreted to represent a portion of lower crust during Permian time (for a review of the Ivrea-Verbano zone see Handy 1987; Handy and Zingg 1991; Handy et al. 1999). This lower crust was overprinted by a retrograde amphibole to greenschist facies mylonite and cataclasite referred to as the Pogallo shear zone (Fountain 1976; Rutter et al. 1999; Handy 1987). Along the Pogallo shear zone upper crustal rocks of the Strona-Ceneri Zone were juxtaposed against lower crustal rocks of the Ivrea zone. Therefore, and because of the ages obtained from this shear zone, Handy et al. (1999) interpreted this shear zone as being responsible for the exhumation of the lower crust to depths of 10 km or less during Jurassic rifting. Furthermore, evidence for exhumation of lower crust to the seafloor is also documented in syn-rift sediments in the Canavese zone, where clasts of migmatitic gneisses and mafic granulites are found in tectonosedimentary breccia (Ferrando et al. 2004). The units of the Adriatic margin exposed in the Southern Alps display a stratigraphic and structural evolution during the Jurassic rifting similar to the Austrolapine nappes further to the northeast (Manatschal 2004; Berra et al. 2009).

3. 3. Preserved relics of the necking zone in the Eastern and Southern Alps sections

In both the Eastern and Southern Alps, earlier work focused mainly on the most proximal and distal parts of the rift sections while the units that are located inbetween are either not well exposed (Gozzano high in the Southern Alps) or have only been little investigated (e.g. Bernina, Campo and Grosina). In this paper we will focus on the Bernina, Campo and Grosina units, which preserve relics of the necking zone of the Adriatic margin (Figs. 4 and 5). We will discuss the major pre-Alpine structures identified in these units based on detailed field mapping, structural, petrological and thermochronological studies.

Fig. III. 3: (a) Tectonic map of the Alps showing the distribution of the major palaeogeographic domains (modified after Schmid et al. 2004). (b) Geological overview of the Eastern and Southern Alps modified after Bernoulli et al. (1990) and Manatschal and Bernoulli (1999). (c) Paleogeographic situation of the Alpine Tethys Ocean and adjacent margins during Early Cretaceous time. (d) Reconstructed palaeogeographic sections across the Alpine Tethys margins, European transect (A-A`) and Eastern Alps transect (B-B') after Mohn et al (subm), Southern Alps transect (C-C`) modified after Bertotti et al. 1993 (for traces of sections see maps above).

4. Alpine overprint in the Margna-Bernina-Campo-Grosina units

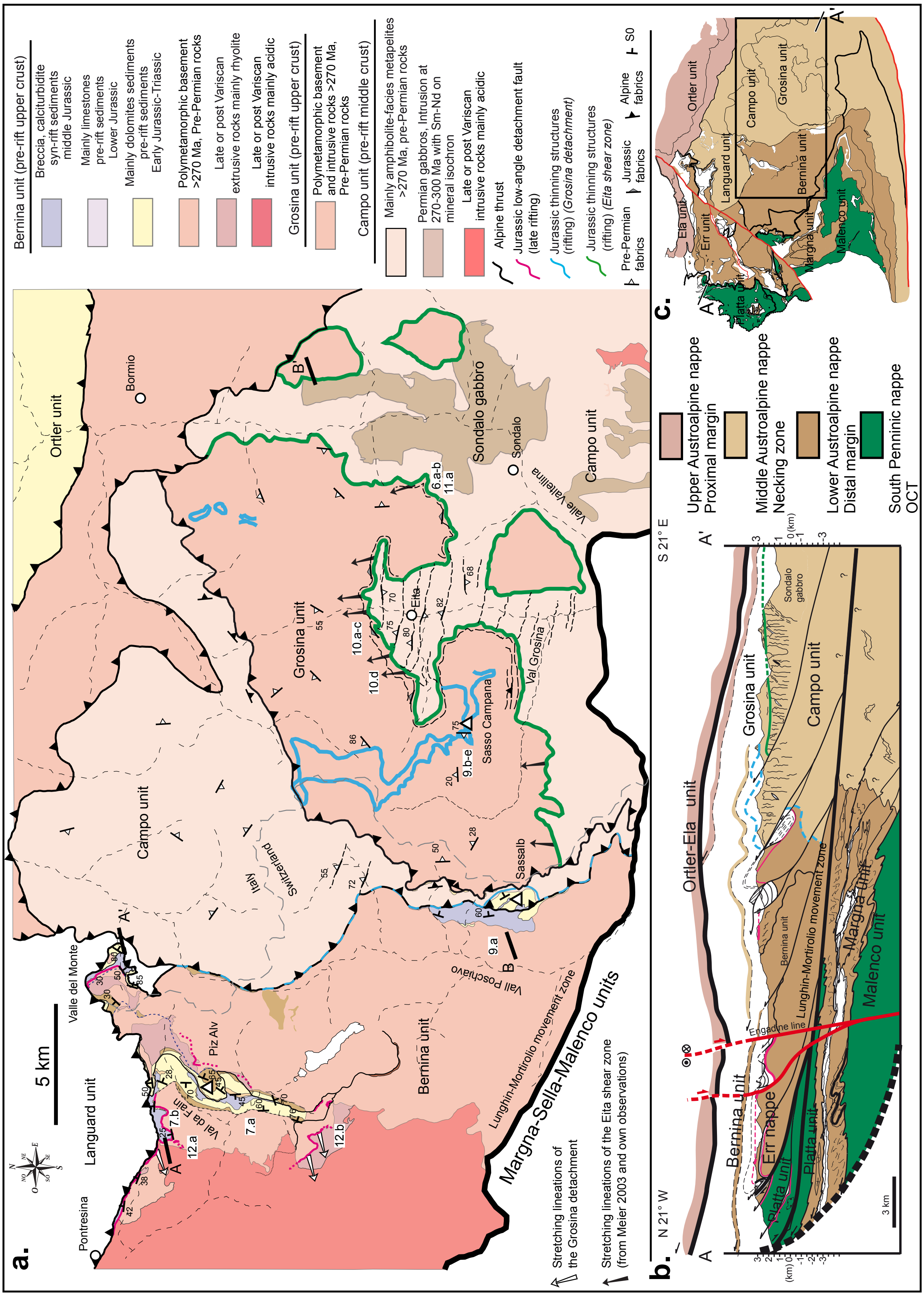
The major Alpine structures observed in Autroalpine nappes and in particular in the Bernina-Campo-Grosina units are Late Cretaceous W to NW directed thrust faults (Fig. 4). These structures correspond to localized shear zones mainly reactivating pre-Alpine structures or localizing within the post-rift sediments. They can be mapped over tens of kilometers defining a nappe stack that consists of basement units with locally preserved slices of pre-rift sedimentary cover. The Alpine metamorphic overprint in these units did not exceed lowermost greenschist facies (Handy et al. 1996). Large-scale, E-W striking folds with subvertical fold axial planes overprint this nappe stack. Although the first order Alpine tectonic and metamorphic overprint is rather simple showing structures similar to those found in present-day fold and thrust belts, in detail the structures are more complex, due to the strong pre-Alpine rift inheritance (Mohn et al subm). It is important to note that the different units forming the Middle and Lower Austroalpine nappe stack are formed by lithologies that at the end of the Permian were located in upper, middle and lower crustal levels. However, each of these units preserves primary contacts to either pre-, syn- or post-rift Mesozoic sediments. This indicates that the Late Cretaceous nappe stack sampled only units that were within the uppermost part of the crustal section (< 5km). Thus, exhumation of various crustal levels had to occur before onset of Alpine deformation but after Permian time, showing that all these units had to undergo major crustal thinning and exhumation during Jurassic rifting. While in a previous paper we focused mainly to the Alpine history (Mohn et al. subm), the aim of this paper is to investigate and discuss the structures responsible for this major crustal thinning. However, these structures are often reworked in Alpine times, we first discuss the Alpine overprint, the nature and origin of the basement rocks, and the stratigraphic record, before focusing on the description of the structures responsible for extreme crustal thinning.

4. 1. Alpine structures and evolution of the Margna- Bernina-Campo-Grosina units

Although all units in the Austroalpine nappe stack in SE-Switzerland and N-Italy were affected by W to NW directed thrusting during the Late Cretaceous, this deformation event is differently expressed in the units from the proximal margin (more internal units), made essentially of sediment units, and those from the distal margin and OCT (more external units), made essentially of basement rocks. The Alpine structures of the Bernina, Campo and Grosina units are strongly controlled by the former architecture of the rifted margin as shown by Mohn et al (subm). In these units, Alpine deformation is restricted to few, well-localized top to the NW thrust faults that are associated with duplex structures and NW facing anti-and synformal fault bend folds with NE-SW trending axes (Fig. 4).

A key area to study the pre-Alpine geometry and its Alpine overprint is the contact between the Bernina and the Campo-Grosina units exposed in Val Poschiavo at the Swiss Italian border (Figs. 4 and 5). Staub (1946) proposed that in this area the contact between the overlying Grosina and the underlying Campo unit consisted of a thrust fault (e.g. Grosina thrust in Fig. 5b), based on the occurrence of a thin slice of Triassic dolomite at the contact. Further to the SE, this thrust truncates an older structure along which the Grosina and Campo units were juxtaposed, hereafter referred to as the Eita shear zone (Meier 2003). This shear zone is interpreted to represent a pre-Alpine structure (see discussion below). Deformation along the Grosina thrust is localized along a narrow, lower greenschist facies shear zone (~ m thick) with stretching lineations defined by K-feldspar porpyroclasts with asymmetric strain fringes indicating a top to the west sense of shear. Based on the cross cutting relationship, geometry (basement above Mesozoic dolomites) and sense of shear, the Grosina thrust can be interpreted as a Late Cretaceous Alpine thrust fault (Staub 1946). The underlying Campo unit is thrust over the Bernina unit along a second thrust fault (refer as the Campo thrust in Fig. 5). This is best observed at Sassalb, due to the presence of relics of the Mesozoic cover. The Mesozoic cover of the Bernina unit in this area is forming a W-facing synform, which is truncated by W-directed splays of the Campo thrust resulting in a complex, imbricate structure (for details see Mohn et al. subm). Although complex, at Sassalb, the overprint between Alpine thrust structures with moderate displacement can be observed to overprint pre-Alpine structures (Eita shear zone).

Similar relationships between Alpine and pre-Alpine structures can also be observed in the Val dal Fain section shown in Fig. 5a, where thrust faults nucleate along a former rift-related detachment fault, forming imbricate and/or duplex structures, which are further complicated by the existence of extensional allochthons and/or the existence of rift-related high-angle faults (for details see Mohn et al. subm).


a.
b.
c.
Bernina unit (pre-rift upper crust)
Breccia, calciturbidite syn-rift sediments middle Jurassic
Mainly limestones pre-rift sediments Lower Jurassic
Mainly dolomites sediments pre-rift sediments Early Jurassic-Triassic
Polymetamorphic basement >270 Ma, Pre-Permian rocks
Late or post Variscan extrusive rocks mainly rhyolite
Late or post Variscan intrusive rocks mainly acidic
Grosina unit (pre-rift upper crust)
Polymetamorphic basement and intrusive rocks >270 Ma, Pre-Permian rocks
Campo unit (pre-rift middle crust)
Mainly amphibolite-facies metapelites >270 Ma, pre-Permian rocks
Permian gabbros. Intrusion at 270-300 Ma with Sm-Nd on mineral isochron
Late or post Variscan intrusive rocks mainly acidic
Alpine thrust
Jurassic low-angle detachment fault (late rifting)
Jurassic thinning structures (rifting) (Grosina detachment)
Jurassic thinning structures (rifting) (Eita shear zone)
Pre-Permian fabrics
Jurassic fabrics
Alpine fabrics
S0
Stretching lineations of the Grosina detachment
Stretching lineations of the Eita shear zone (from Meier 2003 and own observations)
5 km
Pontresina
Languard unit
Valle del Monte
Campo unit
Ortler unit
Bormio
Grosina unit
Sondalo gabbro
Sondalo
Campo unit
Valle Valtellina
Val Grosina
Sasso Campana
Sassalb
Eita
Italy
Switzerland
Piz Alv
Val da Fain
Bernina unit
Val Poschiavo
Lunghin-Mortirolio movement zone
Margna-Sella-Malenco units
Upper Austroalpine nappe
Proximal margin
Middle Austroalpine nappe
Necking zone
Lower Austroalpine nappe
Distal margin
South Penninic nappe
OCT
N 21° W
S 21° E
Ortler-Ela unit
Grosina unit
Campo unit
Sondalo gabbro
Bernina unit
Err nappe
Platta unit
Margna unit
Malenco unit
Engadine line
3 km
Ortler unit
Ela unit
Err unit
Languard unit
Campo unit
Grosina unit
Bernina unit
Margna unit
Malenco unit
Platta unit

In contrast to the Bernina-Campo-Grosina units, which show a weak Alpine overprint, the Margna-Malenco units located to the southwest and separated from the former units by the Lunghin-Mortirolio movement zone are more strongly affected by Alpine deformation (Mohn et al. subm) (Fig. 4). These units show a higher Alpine metamorphic overprint and more distributed deformation that includes also an extensional overprint, which makes that it is more difficult to restore the pre-Alpine deformation history within these units.

Apart from the Late Cretaceous compression, deformation during Tertiary continental collision between Europe and Adria was manifested in the Austroalpine and South Penninic nappes by a N-S compressional event. However, do to the position of the Bernina-Campo and Grosina units between north and south-vergent units in the neutral position of the collisional orogen, this event resulted only in a weak overprint, which is mainly related to large amplitude (~km) folding with E-W trending axes and sub-ordinate north and south vergent thrusts with moderate displacement (< 100m).

Based on a structural investigation and detailed mapping and thanks to the weak Alpine overprint, it is possible to reconstruct the pre-Alpine geometry and the paleogeographic position of the Bernina-Campo-Grosina units (Fig. 5). In this reconstruction, the Bernina unit can be placed oceanwards of the Campo-Grosina units separated by a major structure referred to as the Grosina detachment. In the following section, we will describe the pre-Alpine structure of the Campo-Grosina and the Bernina domains.

5. Pre-Alpine structures

5. 1. Pre-Alpine structures in the Campo-Grosina domain

5. 1. 1. Field relationships and lithologies

The Campo-Grosina units preserve a 3 km thick crustal section that is exposed to the south and east of the Swiss-Italian border (Figs. 4, 5). The Campo unit is separated from the overlying Grosina unit along the Eita shear zone, which is very well exposed in Valle Grosina (Fig. 4). As discussed below, this shear zone is interpreted to be a Jurassic shear zone.

Fig. III. 4: (a) Geological map of the Bernina-Campo-Grosina units in SE-Switzerland and N-Italy. Map compiled from Staub (1946), Bonsignore et al. (1969); Montrasio et al. (1969); Meier (2003) and own observations. (b) Reconstructed Alpine section across the Upper Penninic and Austroalpine nappes in SE Switzerland and N-Italy (c) Tectonic map of the Austroalpine and Upper Penninic units in SE Switzerland and N-Italy.

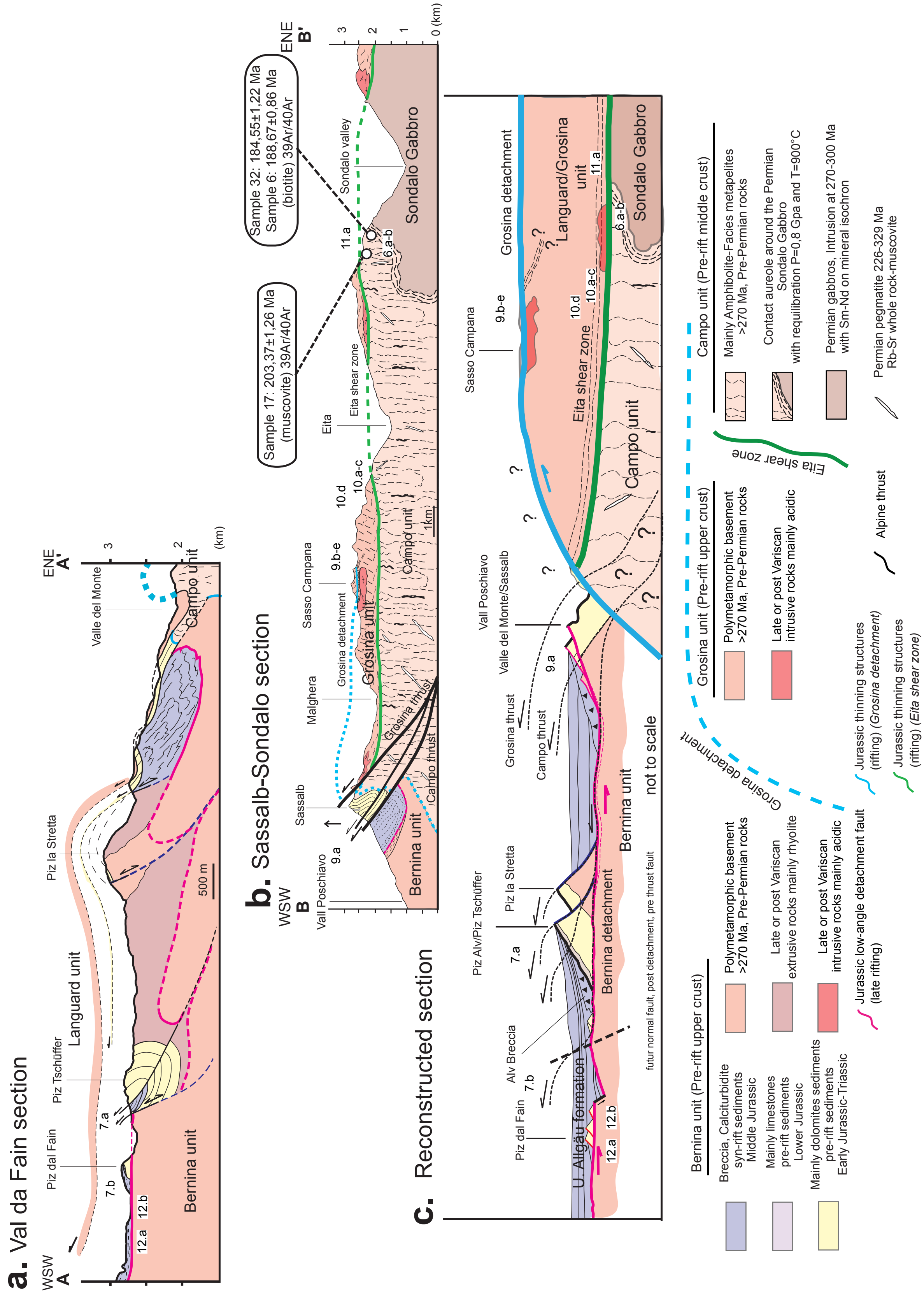


Fig. III. 5: Constructed sections across (a) Val dal Fain, and (b) Sassalb –Sondalo, and (c) restored section showing the pre-Alpine structures and their relation to the sediments and basement rocks in the study area (modified after Mohn et al, subm).

The Grosina unit is exclusively made by Paleozoic basement rocks, which have undergone several phases of pre-Alpine deformation. The Grosina basement is characterized by amphibolite facies metamorphic assemblages, as indicated by the sillimanite-biotite gneiss associated with garnet-biotite-staurolite micaschists, minor diopside-bearing calc-silicates, and variably deformed orthogneiss (Staub 1946; Koeinig 1964; Schudel 1965, Meier 2003). Higher temperature assemblages can be found in this unit indicated by andalusit bearing micaschists. Locally this basement is intruded by granodiorit rocks that cross-cut the previous amphibolites facies rocks. Similar rocks have been described from the adjacent Bernina unit where they are linked with rhyolites (Spillmann and Büchi 1993) dated as late to post Variscan. Therefore we postulate that the granodiorits of the Bernina and Grosina units were emplaced simultaneously in a shallow crustal level.

The Campo unit consists mainly of metapelites, locally associated with amphibolites and calc-silicates, which show a multiphase pre-Alpine tectonic history. In this study, we will mainly focus on the Campo basement exposed to the south of the Grosina unit, in the Valtellina valley (Fig. 4). In this area, the most common metamorphic fabric visible in the field is generally a steeply dipping, E-W striking foliation (Fig. 4a and Fig. 5b). In the micaschists, this fabric is defined by a foliation containing muscovite-biotite-plagioclase-quartz±garnet±staurolite±silliminate, which formed at temperatures between 550 and 650°C (Bucher and Frey 1994; Meier 2003). Similar metamorphic assemblages have been reported by Notarpietro and Gorla (1981), Braga et al. (2001) and Meier (2003) indicating a regional metamorphic equilibration under amphibolite facies condition. Locally the association kyanite+staurolite is found, possibly predating the sillimanite-bearing fabrics (Braga et al. 2001). The absolute timing of formation of the sillimanite-bearing foliation of the Campo unit is unknown. However, similar metamorphic conditions were reached in the adjacent Alpine tectonic units (Oetztal unit and from SouthAlpine basement), at ca. 320-340 Ma (Boriani and Villa 1997; Thöni 1999).

The amphibolite-facies Campo basement in the Valtellina valley underwent contact metamorphism during the intrusion of the Sondalo gabbroic complex, dated at ca. 270-300 Ma (Tribuzio et al. 1999). Intrusion of this mafic complex has been interpreted as related to a major post-Varsican tectonic event, leading to the intrusion of gabbroic rocks at middle to lower crustal levels throughout Western Europe (Rebay and Spalla 2001). Tribuzio et al. (1999) and Braga et al. (2001) initially proposed that the Sondalo gabbro was emplaced at P=0.3-0.8 Gpa. Such estimates have been refined recently by the discovery of prismatine within granulite facies rocks in the contact aureole indicating that re-equilibration occurred at P=0,8 Gpa and T=900°C

(Braga et al. 2003). This enables to propose a middle to lower crustal position of the Sondalo Gabbro and of the surrounding host rocks belonging to the Campo unit during Permian time. The Campo basement was also intruded by acid magmas, leading to the formation of granite, granodiorite and pegmatite, which are all discordant with respect to the pre-Permian Campo fabrics. Acid magmatism has been dated by Rb-Sr method on muscovite-whole rock. Granitoids yielded ages between 282±4 Ma to 259±4 Ma (Del Moro and Notarpietro 1987; Meier 2003 and references therein) while pegmatites are showing a wider age range between 226 to 329 Ma (Hanson et al. 1966; Thöni 1981). More recently Sölva et al. (2003) dated a magmatic garnet from a pegmatite with the Sm/Nd method, for which they obtained crystallization ages of 255,4±2,8 Ma and 250±2,7 Ma from the core and the rim respectively. This poly-phase magmatic evolution is coherent with similar observations from other Alpine basement terrains, which preserve abundant evidence of a major pre-rifting Permian tectonic event (Rebay and Spalla 2001; Schüster and Stüwe 2008 and references therein). In the adjacent Malenco-Margna units (Figs. 4b and c), Hermann et al. (1997) described a Permian gabbro welding the contact between subcontinental mantle and lower continental crust. Müntener et al. (2000) showed, based on detailed petrological investigations and dating, a nearly isobaric cooling for this Permian fossil crust-mantle boundary from Permian to Triassic time. This shows that no significant exhumation occurred during this period, which is also compatible with the occurrence of a wide spread shallow water carbonate platform during the Triassic time. Müntener et al. (2000) were able to determine the P-T conditions at the base of the crust at the onset of rifting in Late Triassic time at 600±50°C at 0,8±0,1 GPa indicating a equilibrated geothermal gradient for the continental crust. The age of exhumation and extreme crustal thinning of these lower crustal rocks are limited to the Jurassic, as indicated by the occurrence of a continuous Permo-Triassic sedimentary sequence evolving from fluviatil silisiclastic sediments to shallow marine carbonates. The above considerations suggest that Permian magmatism and Jurassic rifting are related to two independent tectonic events, separated by a period of relative tectonic quiescence in the study area. Therefore, it is reasonable to think that before Jurassic rifting the Campo basement was still at mid-crustal depth.

5. 1. 2. $^{40}Ar/^{39}Ar$ geochronology of the Campo basement

In order to constrain the tectono-thermal evolution and exhumation history of the Campo basement, two samples were collected from near the Sondalo intrusive. The emplacement of the Sondalo Gabbro during Permian time in amphibolites facies metapelites country rocks (Braga et al. 2001) at a mid-crustal level enables to constrain the position of the Campo unit at that time. A contact metamorphic overprint of the country rock can be observed within the 20 to 30 m thick aureole, which preserves abundant evidence of partial melting of the metapelites with extraction of peraluminous anatectic liquid associated with crystallization of granitoid pods (Braga et al. 2001). Locally, migmatitic banding developed concordant with the regional, steeply dipping foliation of the metapelitic rocks. Commonly the leucosome is largely quartzo-feldspathic and garnet is also found, indicating melting of the metapelite protolith at temperatures close to 850-900°C (Patiño Douce and Johnston 1991). The melanosome is rich in biotite and sillimanite and define an anastomosing fabric. Migmatization is generally restricted to narrow, few tens of meters thick belts surrounding the intrusion. Associated with the gabbro intrusion are also granitoid dykes that contain blocks of refractory granulite xenolith (Braga et al. 2001). Samples for $^{40}Ar/^{39}Ar$ geochronology were taken from the migmatites and quartzo-feldspathic dykes within the contact metamorphic aureole of the Sondalo Gabbro. Localization of sample and descriptions are given in table 1, step-heating data are listed in table 2.

Sample	Host rock	Petrography of sample
Sample 32, 822096/138665 (Swiss topographic coordinates)	Banded migmatite at about 30 meters from the gabbro-country rock contact.	● Pervasive foliation defined by the preferred orientation of white mica+biotite in equilibrium with quartz+garnet+plagioclase ● Local retrogression of biotite to chlorite is observed ● The biotite-bearing foliation is also present within quartz porphyroblasts, indicating that grain size coarsening of the quartzitic matrix post-dated deformation.
Sample 6, 822574/138236 (Swiss topographic coordinates)	Leucocratic layer of a migmatite characterized by compositional banding.	● Restit: thin, spaced layers made of fine-grained aggregates of sillimanite+biotite I+quartz define a weak anastomozing foliation concordant with the foliation in the metapelitic country rocks. ● Leucosome: porphyroblastic garnet is found with inclusions of biotite and quartz. Garnets are associated with a second generation of large biotite II+coarse anhedral quartz+subhedral plagioclase ● A younger generation of fine-grained biotite III is present in cracks of the garnet. Three different generations of biotite could be identified in this rock based on microstructural observations.

Table 1: Samples locations (with Swiss grid coordinates) and descriptions of the mineral assemblages.

Laser	40Ar/39Ar	37Ar/39Ar	36Ar/39Ar	40Ar*/39Ar$_k$	40Ar(mol)	40Ar* (%)	39Ar$_k$ (%)	K/Ca	Age	±2σ
Power (w)										
Sample 32 (biotite)										
1,4	17,28995	0,00002	0,02851	8,863757	1,837E-16	51,27	0,31	100,769	97,65	30,26
1,7	16,96400	0,00002	0,00251	16,221879	5,066E-15	95,63	8,66	93,585	174,88	0,93
2,0	17,15386	0,00285	0,00108	16,834160	3,878E-15	98,14	6,55	150,914	181,16	1,77
2,2	17,18308	0,00327	0,00082	16,938737	4,007E-15	98,58	6,76	131,410	182,23	1,19
2,6	17,12051	0,00558	0,00050	16,973188	4,999E-15	99,14	8,46	77,111	182,59	0,86
2,9	17,26758	0,00537	0,00001	17,266628	4,713E-15	99,99	7,91	80,103	185,59	0,21
3,1	17,29408	0,00533	0,00026	17,216628	4,984E-15	99,55	8,35	80,641	185,08	0,92
3,5	17,32040	0,00665	0,00002	17,319463	3,902E-15	99,99	6,53	64,652	186,13	0,21
3,9	17,26482	0,00312	0,00002	17,263845	4,333E-15	99,99	7,27	137,886	185,56	0,25
4,3	16,99212	0,00369	0,00001	16,991153	3,980E-15	99,99	6,79	116,468	182,77	0,25
5,1	17,04185	0,00370	0,00009	17,013126	4,328E-15	99,83	7,36	116,103	183,00	1,22
6,0	16,95342	0,00414	0,00010	16,921916	4,836E-15	99,81	8,27	103,847	182,06	1,01
7,2	16,85759	0,00001	0,00001	16,856582	2,372E-15	99,99	4,08		181,39	0,30
8,5	17,17520	0,00535	0,00030	17,086371	3,397E-15	99,48	5,73	80,447	183,74	1,25
10,4	16,85288	0,01143	0,00027	16,771988	4,044E-15	99,52	6,96	37,614	180,53	1,43
sample 6 (biotite)										
1,4	19,92750	0,00747	0,01966	14,117745	9,265E-16	70,85	0,28	57,599	153,02	6,92
1,7	18,60404	0,00903	0,00696	16,547378	1,707E-15	88,94	0,55	47,637	178,09	3,27
2,0	18,01731	0,00023	0,00394	16,852152	3,368E-15	93,53	1,11	1890,963	181,21	1,63
2,2	17,95177	0,00010	0,00199	17,361803	7,996E-15	96,71	2,65	46,595	186,42	0,98
2,6	17,93330	0,00010	0,00093	17,658264	1,794E-14	98,47	5,96	46,595	189,44	0,33
2,9	17,66111	0,00010	0,00091	17,391271	8,382E-15	98,47	2,83	46,595	186,72	0,59
3,1	17,20395	0,00010	0,00036	17,097192	9,218E-15	99,38	3,19	46,595	183,72	0,63
3,5	17,56944	0,00010	0,00076	17,343906	1,192E-14	98,72	4,04	46,595	186,23	0,53
3,9	17,84099	0,00010	0,00059	17,666325	1,398E-14	99,02	4,67	46,595	189,52	0,52
4,3	17,85474	0,00010	0,00036	17,746593	1,898E-14	99,39	6,33	46,595	190,34	0,44
5,1	17,08047	0,00010	0,00008	17,054481	2,479E-14	99,85	8,64	46,595	183,28	0,31
6,0	17,67840	0,00015	0,00010	17,647339	3,272E-14	99,82	11,02	46,595	189,33	0,27
7,2	17,57790	0,00015	0,00015	17,576890	4,151E-14	99,99	14,06	46,595	188,61	0,19
8,5	17,55766	0,00015	0,00015	17,556655	3,831E-14	99,99	12,99	46,595	188,40	0,19
10,4	17,83295	0,00015	0,00015	17,831936	6,496E-14	99,99	21,69	46,595	191,21	0,18
Sample 17 (muscovite)										
1,4	23,53434	0,00901	0,02792	15,283248	9,711E-16	64,94	0,38	47,718	165,04	7,35
1,7	19,81151	0,00064	0,00441	18,506146	6,116E-15	93,41	2,83	677,038	197,99	0,75
2,0	19,82142	0,00001	0,00366	18,738982	8,449E-15	94,54	3,91	1555,327	200,35	0,68
2,2	19,36670	0,00001	0,00249	18,630308	1,688E-14	96,20	7,99	1555,327	199,25	0,39
2,7	19,93616	0,00001	0,00285	19,093693	1,382E-14	95,77	6,35	1555,327	203,94	0,42
3,1	19,53512	0,00001	0,00199	18,945903	8,582E-15	96,98	4,03	1555,327	202,44	0,55
3,4	19,97448	0,00001	0,00231	19,289609	1,475E-14	96,57	6,77	1555,327	205,91	0,39
4,2	19,92824	0,00001	0,00278	19,104887	1,818E-14	95,87	8,36	1555,327	204,05	0,42
4,8	19,93046	0,00001	0,00296	19,053345	1,950E-14	95,60	8,97	1555,327	203,53	0,41
5,6	19,60944	0,00001	0,00207	18,996279	2,820E-14	96,87	13,18	1555,327	202,95	0,31
6,5	19,16455	0,00001	0,00105	18,852912	2,825E-14	98,37	13,51	1555,327	201,50	0,35
8,8	19,14233	0,00001	0,00049	18,997759	2,069E-14	99,24	9,91	1555,327	202,97	0,35
10,4	19,29823	0,00001	0,00038	19,186045	2,914E-14	99,42	13,84	1555,327	204,87	0,29

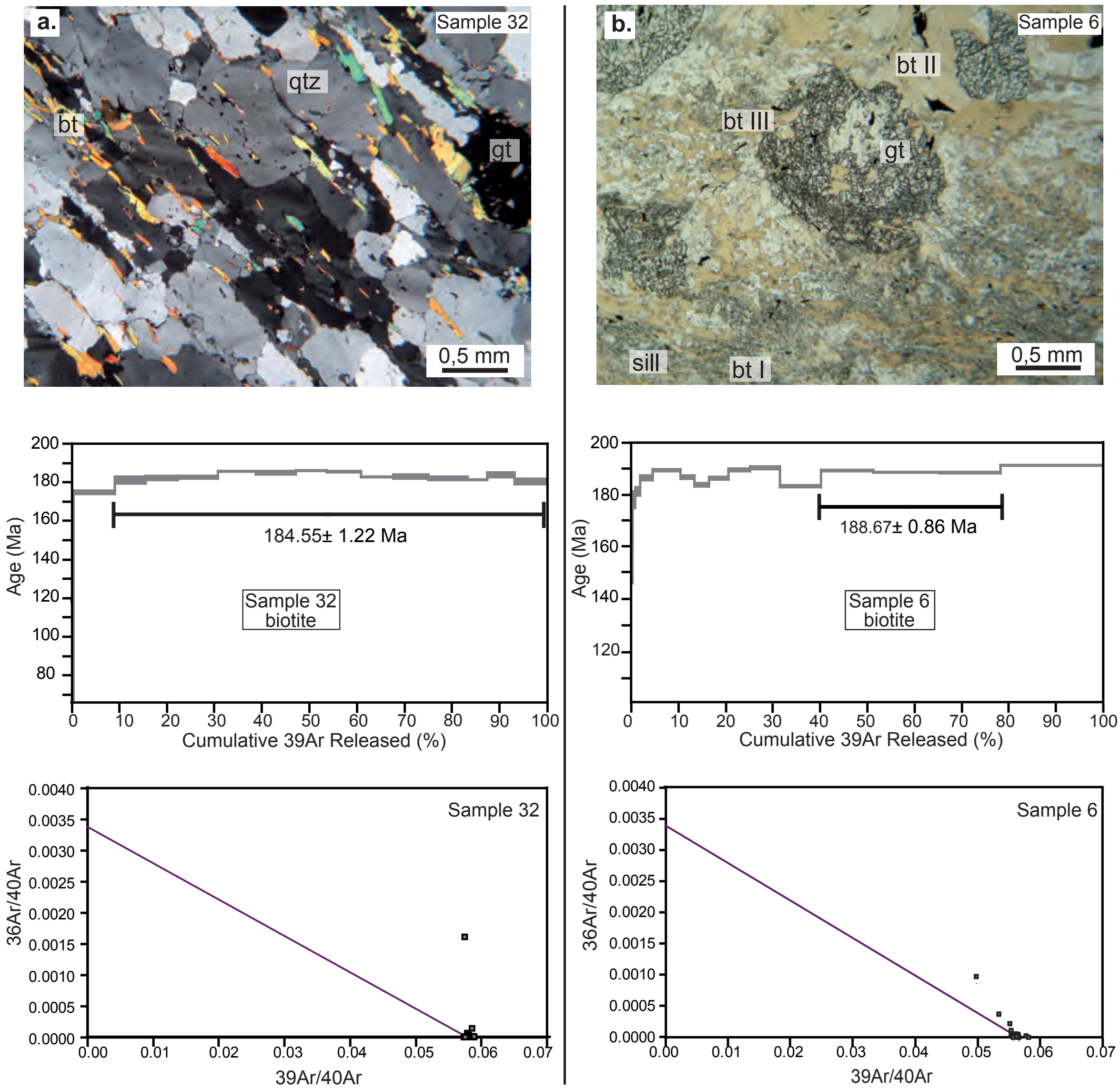
Table 2: $^{40}Ar/^{39}Ar$ step-heating data.

5. 1. 3. Analytical techniques

Samples were crushed and sieved and biotite and white mica were hand picked from the fraction ranging between 250 and 500 µm. The samples were irradiated for 24 hours in the OSU, CLICIT facility, and J values were calculated via the irradiation of FCT SAN (28.02 +/-0.16 Ma; Renne 1994) samples, which were separated by distances of less than 1 cm throughout the columnar irradiation package. 39Ar-40Ar analyses were performed at the laboratory at the University of Geneva with an Argus (GV Inst.) multi-collector mass spectrometer, equipped with four high-gain (10E12 ohm) Faraday collectors for the analysis of 39Ar, 38Ar, 37Ar and 36Ar as well as a single Faraday collector (10E11 ohm) for the analysis of 40Ar. The automated UHV stainless steel gas extraction line incorporates one SAES AP10 getter, and one SAES GP50-ST707 getter. White mica and biotite were step-heated using a defocused 30 W, MIR10 IR (CO2) laser that was rastered over the samples to provide even heating of the grains. Samples were measured on the Faraday collectors and time-zero regressions were fitted to data collected from 12 cycles. Peak heights and blanks were corrected for mass discrimination, isotopic decay of 39Ar and 37Ar, and interfering nucleogenic Ca-, K- and Cl-derived isotopes. Baselines were measured prior to each step analysis. Error calculations include the errors on mass discrimination measurement and the J value. 40Ar, 39Ar, 38Ar, 37Ar and 36Ar blanks were calculated before every new sample and after every three heating steps. 40Ar blanks were between 6.5E-16 and 1.0E -15 moles. Blank values for m / e 39 to 36 were all less than 6.5E -17 moles. Age plateaux were determined using the decay constant of Steiger and Jaeger (1977) and the criteria of Dalrymple and Lanphere (1974). The automated analytical process uses the software ArArCalc (Koppers, 2002). The samples were irradiated for 24 hours in the OSU, CLICIT facility, and J values were calculated via the irradiation of FCT SAN (28.02 +/-0.16 Ma; Renne 1994) samples, which were separated by distances of less than 1 cm throughout the columnar irradiation package.

5. 1. 4. Results of $^{40}Ar/^{39}Ar$ step-heating of biotite (Fig. 6)

Sample 32 yielded a relatively flat apparent age spectrum, starting from an age of 174 Ma, increasing progressively to 186 Ma. A ”plateau” age of 184.55±1,22 Ma can be calculated (Fig. 6). The distribution of data points in the $^{39}Ar/^{40}Ar$ vs $^{36}Ar/^{40}Ar$ plot does not allow to test whether any inherited argon component is present in the dated biotite (McDougall and Harrison 1988). As shown above petrographic and microstructural observations indicate the presence of a single generation of biotite. This single generation of biotite is stable with garnet and plagioclase in the Permian fabric related to the Sondalo intrusion. The presence of a single gas reservoir can also be deduced by the flat age spectrum. Therefore, the age of 184.55±1.22 Ma can be interpreted as a cooling age.

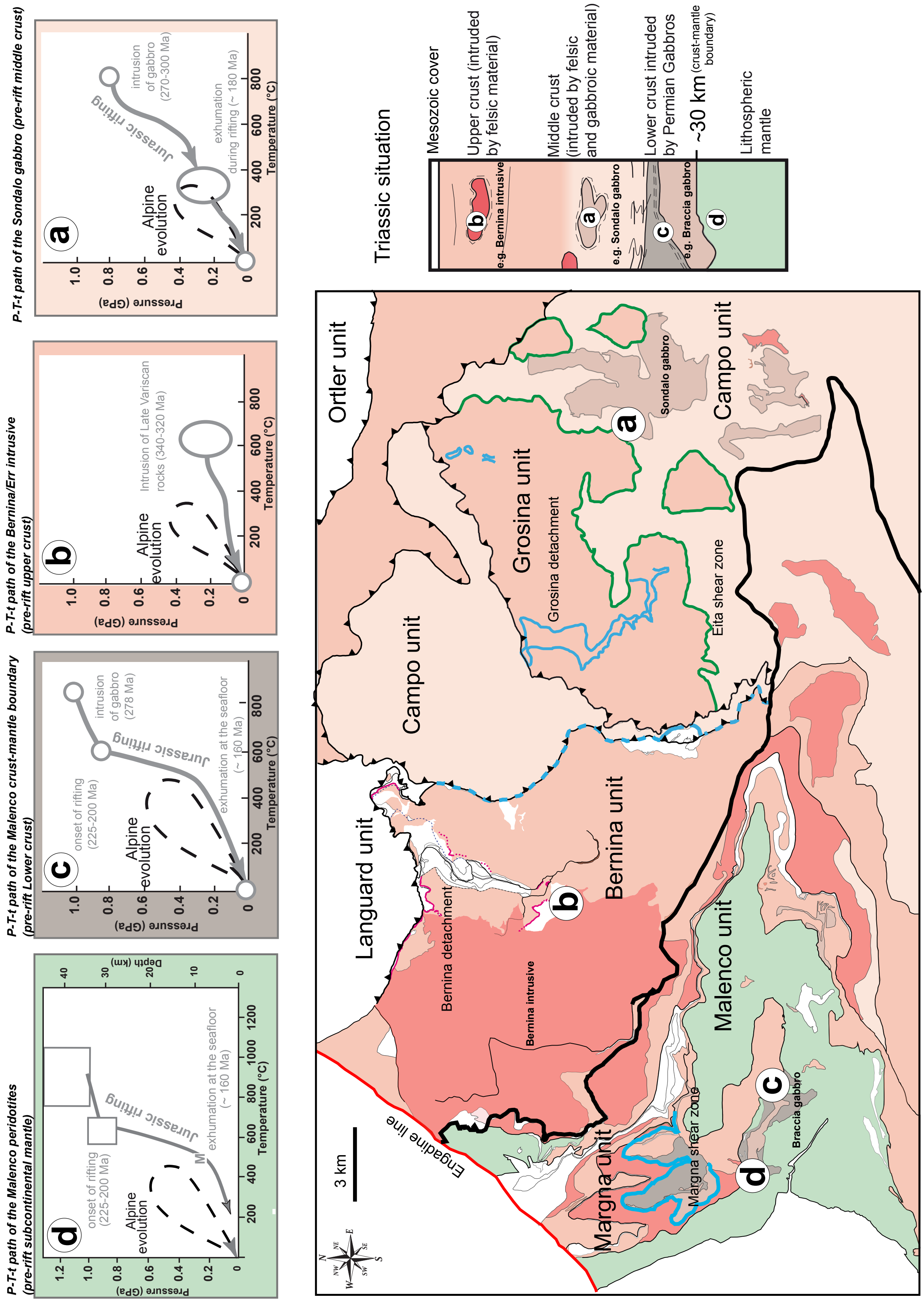


The apparent age spectrum of sample 6 is more complex. The first steps yielded apparent ages in the 178-189 Ma range and then evolve in a flat apparent age spectrum of 188.67±0,86 Ma (Fig. 6). The $^{39}Ar/^{40}Ar$ vs $^{36}Ar/^{40}Ar$ plot gives no significant information, due to the clustering of data points next to the $^{39}Ar/^{40}Ar$ axis. Based on microstructural data, three different generations of biotite could be identified in this sample. Biotite I is stable in the melanosome and probably inherited from a pre-Permian fabric, whereas biotite II is found in the leucosome in equilibrium with poryprobasltic garnet in relation with the metamorphic aureole of the Permian Sondalo gabbro. Furthermore, a third generation of biotite is observed. Each generation of biotite is characterized by a distinct grain size. Therefore, the apparent age spectrum is likely to be the result of mixing of different argon reservoirs. The relatively minor age spread may be explained as being related to slightly different retentivities of the dated biotites, due to their different grain sizes. In conclusion, the age obtained from this sample is interpreted as a cooling age of the Campo basement.

5. 1. 5. Significance of the measured ages

Biotite cooling ages indicate that, during Permian and Triassic time, the Sondalo gabbro resided in the middle crust, before being exhumed and cooled in the Jurassic (Fig. 7a). Lister et al 1996 have emphasized the possible loss of argon by diffusion during slow cooling in the vicinity of the closure temperature. In our study the shape of the age spectrum are coherent with rapid cooling without significant loss of Argon by diffusion. These results are confirmed by previous studies, which yielded similar ages with $^{40}Ar/^{39}Ar$ on white mica (188±4 Ma, Meier 2003), K/Ar on biotite (157-190 Ma, Thöni 1981), K/Ar on white mica (217±11 Ma, Hanson et al. 1966). Amphibole from dioritic rocks of the Sondalo intrusion have been dated by $^{40}Ar/^{39}Ar$, yielding an age of about 300 Ma (Potenza et al. 1991). $^{40}Ar/^{39}Ar$ ages on biotites are showing the rapid cooling of the Campo basement during early Jurassic time (184 to 188 Ma). Despite uncertainties on the retention properties of biotite, the results suggest that, prior to exhumation, the Campo basement was at temperatures in excess of 350-400°C. These data can be compared

Fig. III. 6: Photomicrographs of the samples from the Campo unit that have been chosen for dating (a) Sample 32: banded migmatite sampled in the vicinity of the Permian Sondalo gabbro (822096/138665 Swiss topographic coordinates) and $^{40}Ar/^{39}Ar$ step-release sepectra and $^{39}Ar/^{40}Ar$ vs $^{36}Ar/^{40}Ar$ plot. (b) Sample 6: migmatite characterized by compositional banding sampled in the vicinity of the Permian Sondalo gabbro (822574/138236 Swiss topographic coordinates) and $^{40}Ar/^{39}Ar$ step-release sepectra and $^{39}Ar/^{40}Ar$ vs $^{36}Ar/^{40}Ar$ plot. (For discussion of the data see text and for data see Table 1, 2).



P-T-t path of the Sondalo gabbro (pre-rift middle crust)
P-T-t path of the Bernina/Err intrusive (pre-rift upper crust)
P-T-t path of the Malenco crust-mantle boundary (pre-rift Lower crust)
P-T-t path of the Malenco peridotites (pre-rift subcontinental mantle)
Pressure (GPa)
Temperature (°C)
Depth (km)
Alpine evolution
Jurassic rifting
intrusion of gabbro (270-300 Ma)
exhumation during rifting (~ 180 Ma)
Intrusion of Late Variscan rocks (340-320 Ma)
onset of rifting (225-200 Ma)
intrusion of gabbro (278 Ma)
exhumation at the seafloor (~ 160 Ma)
Triassic situation
Mesozoic cover
Upper crust (intruded by felsic material)
Middle crust (intruded by felsic and gabbroic material)
Lower crust intruded by Permian Gabbros
~30 km (crust-mantle boundary)
Lithospheric mantle
e.g. Bernina intrusive
e.g. Sondalo gabbro
e.g. Braccia gabbro
Ortler unit
Campo unit
Sondalo gabbro
Grosina unit
Grosina detachment
Eita shear zone
Languard unit
Bernina detachment
Bernina intrusive
Bernina unit
Malenco unit
Margna unit
Margna shear zone
Braccia gabbro
Engadine line
3 km

with results published by Müntener et al. 2000 who have determined that during Triassic time the Adriatic continental crust was thermally equilibrated. Therefore we can postulate that at the beginning of rifting the Campo rocks were at 15 to 25 km depth. As discussed below, we assume that the rapid cooling was related to exhumation along large scale extensional shear zones and detachment faults during Jurassic rifting.

5. 2. Pre-Alpine structures in the Bernina domain

The basement of the Bernina unit is made of Carboniferous to early Permian (333 -295 Ma) calc-alkaline and alkaline igneous rocks (Fig. 7a) (Spillmann 1993; von Quadt et al. 1994) intrusive into poly-metamorphic Variscan basement. The occurrence of Permian rhyolitic bodies formed near or at the surface shows that the Bernina basement was in an upper crustal position already during Permian time (Fig. 7). Mesozoic sediments occur only locally at Piz Alv, Val del Monte and Sassalb slices (Fig. 4a). A key to understand the pre-Alpine evolution of the Bernina unit is the Piz Alv – Val dal Fain area (Fig. 4a and Fig. 8). This area preserves pre-Alpine contacts between Mesozoic sediments and basement rocks, separated by a Jurassic low-angle detachment fault, referred to as the Bernina detachment fault. The overall Alpine structure is well illustrated in the WSW to ENE striking section in Fig. 5a, which is parallel to the main Alpine transport direction. The Piz Alv forms, like Piz Tschüffer, a W-facing recumbent fold that is truncated by a west-vergent second order Late Cretaceous thrust fault that juxtaposes an inverted limb in the hanging wall against a subvertical hinge in the footwall (Fig. 8b). The complex Alpine structure results from the partial reactivation of a former Jurassic detachment fault overlain by an extensional allochthon (see Fig. 5c and Mohn et al. subm for a more detailed description). This detachment fault, referred to as the Bernina detachment, is a major structure that will be discussed in more detailed in a later part of this paper.

At Piz Alv, a complete pre-rift succession is preserved ranging from Permian rhyolites, to massive shallow marine Triassic dolomites to Upper Triassic lagoonal sediments, overlain by Lower Jurassic limestones. The whole sequence is up to 600 to 800 m thick and shows, except for the Lower Jurassic carbonates, a very similar evolution like that observed in the more proximal parts of the Adriatic margin (e.g. Ortler and Ela nappes). Of major importance for the un-

Fig. III. 7: Alpine tectonic map of the study area showing the distribution of the major units discussed in this paper. The section to the right shows a simplified cross-section through the crust in later Permian time, indicating the original position of the rocks found today in the different tectonic units. The P-T-t paths are shown for (a) pre-rift middle crustal gabbros (Sondalo gabbro (Tribuzio et al. 1999; Bragga et al. 2001, 2003); (b) pre-rift upper crustal granitoid (Bernina and Err intrusives (Peters 2005, 2007)); (c) pre-rift lower crustal rocks (Malenco crust – mantle boundary (Müntener et al. 2000; Villa et al. 2000)); (d) pre-rift subcontinental mantle (Malenco and Platta (Müntener et al. 2010 and references therein)). Note that these rocks show evidence for a Jurassic exhumation history.

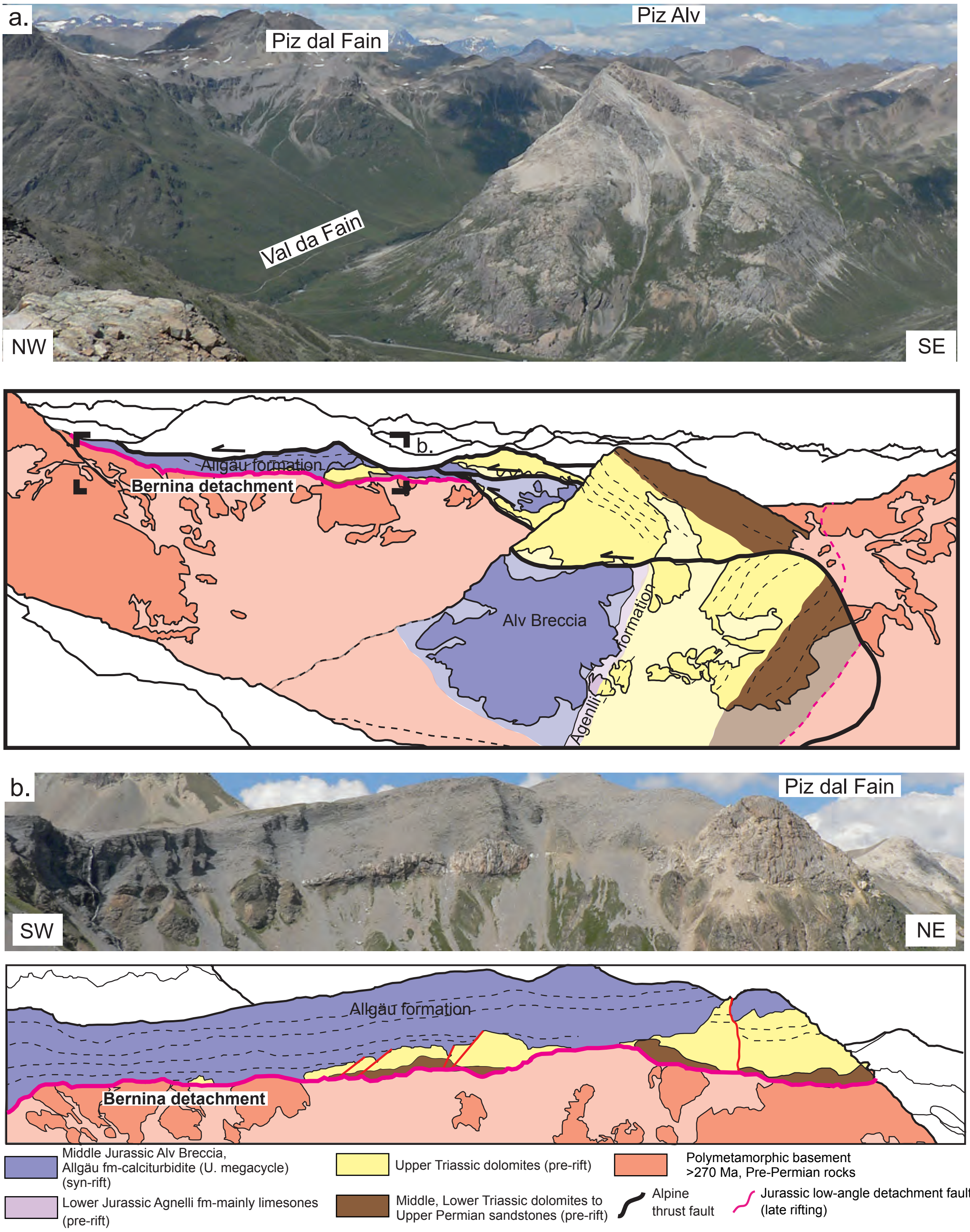


Fig. III. 8: Photograph and line drawing from the Piz Alv – Val dal _Fain area (Bernina domain) (for localization of the photograph see Figs. 4a and 5a) (a) large scale structure of Piz Alv-Val da Fain area view from the S showing the relationship between pre- and syn-rift sediments and basement rocks (796000/148000, Swiss topographic coordinates) (for restoration see Fig. 5c). (b) View of Piz dal Fain from the S showing the syn-rift sediments (Allgäu formation (upper mega-cycle)) onlapping either onto exhumed basement or pre-rift sediments. (795000/149000; Swiss topographic coordinates) (For more details see text).

derstanding of the rift evolution are the Lower Jurassic carbonates that are dated by ammonites as Late Sinemurian to Pleinsbachian in age (Finger 1978; Manatschal and Nievergelt 1997). It highlights that the Bernina domain underwent, like the reminder of the distal margin (e.g. Err domain; Manatschal et al. 2007) a different rift evolution comparing to the more proximal parts of the margin. While the major rift basins in the proximal margin formed during and were filled by thick lower Jurassic syn-rift sediments, the former distal margin, including the Bernina domain became sedimentary starved during this stage and remained shallow. During late Pliensbachian time, deformation migrated into the future distal margin. This is indicated by the sealing of the basins in the proximal margin and the formation of syn-rift sediments including up to 200 to 300 m thick breccias, referred to as the Alv Breccia (Schüpbach 1973) in the distal margin. These breccias consist exclusively of clasts of Triassic and Lower Jurassic carbonate that range in size from the mm to several tens of meters. Previous authors (Schüpbach 1973, Furrer 1985, Froitzheim 1988) interpreted these breccias as formed in situ or by minor gravitational transport at the flanks of submarine highs related to high-angle fault activity. Laterally, the Alv breccias pass into calciturbidites, marls and hemipelagic limestones referred to as Allgäu formation (Eberli 1988). The overall sequence shows a general thinning- and fining upward megacycle interpreted to represent a basinal sequence ranging in thickness between 200 and 500 m deposited in fault-bounded rift basins (Eberli 1988). To the northwest at Piz dal Fain, the syn-rift Allgäu formation is directly overlying fragments of pre-rift sediments (Fig. 8b). Due to the primary nature of the contact with the basement as well as the fact that the Allgäu formation is not deformed at the base, this contact was interpreted as a former high onto which the syn-rift sediments were onlapped (Schüpbach 1973). However, the detailed study of the contact, including the deformation in the basement and the structures in the overlying pre-rift sediments suggest that this structure represents a low-angle detachment (see discussion below). Therefore, these new observations are not in agreement with the previous interpretation suggesting that these sediments were deposited in a simple fault bounded basin.

The sediments of the Allgäu formation observed in Val dal Fain are very similar to those observed in the proximal margin in the Ela and Ortler units. Eberli (1988) showed that the Allgäu formation can be subdivided in the proximal margin in two megacycles, a lower one, Hettangian to Bathonian in age and an upper one late Sinemurian to Bathoninan in age. Eberli (1988) showed that the lower sequence was mainly deposited in a syn-rift setting in the proximal margin, while the upper sequence was interpreted as a post-rift sequence due to the fact that it seals high-angle faults in the proximal margin. In the Bernina domain, the time equivalent unit of the first megasequence is represented by a starved carbonate platform (e.g. Agnelli formation), which shows pre-rift characteristics. The time equivalent unit of the upper sequence in the distal margin corresponds to the sediments exposed in the Alv and Val dal Fain section. These sediments are clearly syn-tectonic. These observations indicate a shift of extension during Pliensbachian to Toarcian time (± 180Ma) from the proximal to the future distal margin.

5. 3. Pre-Alpine structures in the Margna unit

The Margna unit has been described by many authors (Guntli and Liniger 1989; Hermann and Müntener 1996; Hermann et al. 1997; Bissig and Hermann 1999). These previous investigations were able, despite the strong Alpine tectonic overprint, to unravel the pre-Alpine structures in this unit. The most important observation is that the Margna unit (Fig. 7) is formed by two distinct crustal units: an upper one formed by Mesozoic sediments preserving primary contacts to upper crustal rocks including Permian meta-volcanics and a lower one consisting of pre-rift lower crust welded by a Permian gabbro (dated at 270±6 Ma with U-Pb on zircon (Hansmann et al. 1996)) to the subcontinental mantle. Müntener et al. (2000) showed that the magmatic assemblage of the gabbro (e.g. Braccia and Fedoz gabbro) was formed at 9 to 11 kbar at 750-850°C subsequently it underwent an isobaric cooling during the Triassic time until 600±50 °C and 0.8±0,1 Gpa indicating that before Jurassic rifting these rocks were residing at ±30 km depth (Fig. 7c). Hermann and Müntener (1996) and Bissig and Hermann (1999) showed that the contact separating the two units, also referred to as the Margna shear zone, is a pre-Alpine contact. This contact consequently juxtaposes uppermost and lowermost crust, suggesting that more than 20 km of mainly middle crustal rocks were omitted along this contract during Jurassic rifting. The nature and importance of this contact will be discussed in the next section. Apart from the lower crustal rocks forming the lower part of the Margna unit, all remaining basement rocks forming the most distal parts of the former Adriatic margin today exposed in the Lower Austroalpine and South Penninic units are either formed by pre-rift upper crustal rocks or subcontinental mantle rocks (Fig. 7). The lack of mid-crustal rocks in the most distal margin may be a further evidence that the ductile middle crust was omitted in the necking zone further continentwards when the crust was thinning from 30 to 10 kilometers.

6. Pre-Alpine extensional structures accommodating crustal thinning

In the previous sections we showed that the Grosina-Campo units are made of upper and middle crustal rocks while in the Bernina and Margna units located more oceanwards middle crustal rocks are missing (Fig. 7). In the following section, we will focus to the structures that were responsible for crustal thinning. These structures are: (1) The Grosina detachment juxtaposing units belonging to the necking zone (Grosina-Campo) in the footwall from units forming the distal margin (Bernina) in the hanging wall; (2) the Eita shear zone separating pre-rift upper crustal (Grosina) from pre-rift middle crustal (Campo) rocks; (3) the Margna shear zone located in the distal margin and separating pre-rift upper and lower crustal rocks; and (4) the Bernina detachment separating exhumed upper crust from delaminated extensional allochthons and syn- and post rift sediments. Another type of structure that is not preserved in the Middle Austroalpine units, but that may be of importance to explain crustal thinning is the Pogallo shear zone, that is exposed in the Ivrea-Verbano zone (Fig. 3b, d) (for details see discussion).

6. 1. Grosina detachment

The Grosina detachment represents a pre-Alpine brittle fault zone marking the contact between the Bernina and Campo-Grosina units. At Sassalb the primary contact is preserved between the Bernina Triassic dolomites and the Campo basement. As shown in Figs. 5b and 9a, it is folded and cut by the Alpine Campo and Grosina thrust faults. Based on cross-cutting relationships and a restoration (see Fig. 5c) we can demonstrate that the middle crustal rocks of the Campo unit were already juxtaposed against pre-rift dolomites during rifting along an oceanward dipping detachment fault preserving characteristic brittle fault rocks (e.g. green cataclasites and black gouges described by Manatschal 1999 from the Err detachment). We correlate this brittle fault zone with a fault system exposed further to the east where at least two well developed sub-horizontal brittle fault zones can be mapped, spectacularly exposed at the top of the Grosina unit (Fig 5b and 9b). These fault zones belong to a structure that we refer to as the Grosina detachment. The lower brittle fault zone constitutes a continuous, less than 10 meters thick fault zone, which can be followed over a distance of more than 8 km (Fig. 5b). Locally, a second fault zones can be mapped, which is cross cut by the former. This second fault may either represent a branch of a second detachment fault that was cut by the fist one (see incisement structures described by Lister and Davis 1989), or may represent an anastomouzing fault belonging to one and the same detachment system. The detachment faults are made of cataclastic fault rocks showing angular clasts and transgranular fractures passing upwards into phyllosilicate-rich gouge (Fig. 9c). No crystal plastic or dynamic recrystallization of quartz can be observed along the detachment fault, indicating that these fault zones were active at less than 300°C. Stretching lineations are poorly defined and are highly variable, with a predominant E-W direction. A sense of shear remains difficult to constrain due to the lack of shear indicators.

Based on cross-cutting relationship an age for these detachment faults can be proposed. In Val Poschiavo, the detachment fault is truncated by the oldest Alpine thrust faults, suggesting that it is therefore older than the Late Cretaceous Alpine overprint. The observation that along this detachment Permian middle crust rocks (Campo basement) are juxtaposed against Triassic dolomites (Bernina unit) excludes a Permian or older age. This is also in line with the observation that the Grosina detachment cut across Permian high-T fabrics in the Grosina and Campo units. Based on these observations, we propose a Jurassic age for the Grosina detachment.

At the top of the Grosina unit, the detachment fault is locally cross-cutting quartz mylonites. These quartz mylonites can also be observed as clasts in the cataclasites and gouges defining the Grosina detachment (Fig. 9e). The mylonitic shear zone is 2 to 3 m wide and is discordant to the fault zone. The angle between the fault zone and the mylonite shear zone is approximately 20°. The shear zone is discontinuous and showing an anatomizing geometry. The quartz mylonites are formed by quartz+chlorite+illmenite, while plagioclase and K-feldspar and white mica are preserved as porphyroclasts. Fine-grained aggregates of quartz crystals display a strong lattice preferred orientation indicating that strain is largely accommodated through dynamic recrystallization of quartz (Fig. 9d). The evidence of brittle deformation affecting porpyroclastic feldspar constrains the temperature condition for the activity of this shear zone between 300°C and 500°C. Locally, the foliation is also defined by fine aggregates of white micas. A NE-SW stretching lineation defined by elongated K-feldspar porphyroclastes is observed in the quartz mylonite. From shear band and mica fish a top to the NE sense of shear can be inferred for this mylonitic shear zone. Based on field relations, it is not clear whether the quartz mylonite is related to the detachment fault or if it is unrelated to this structure and may represent an older structure. At the moment no age constrains are available for this structure.

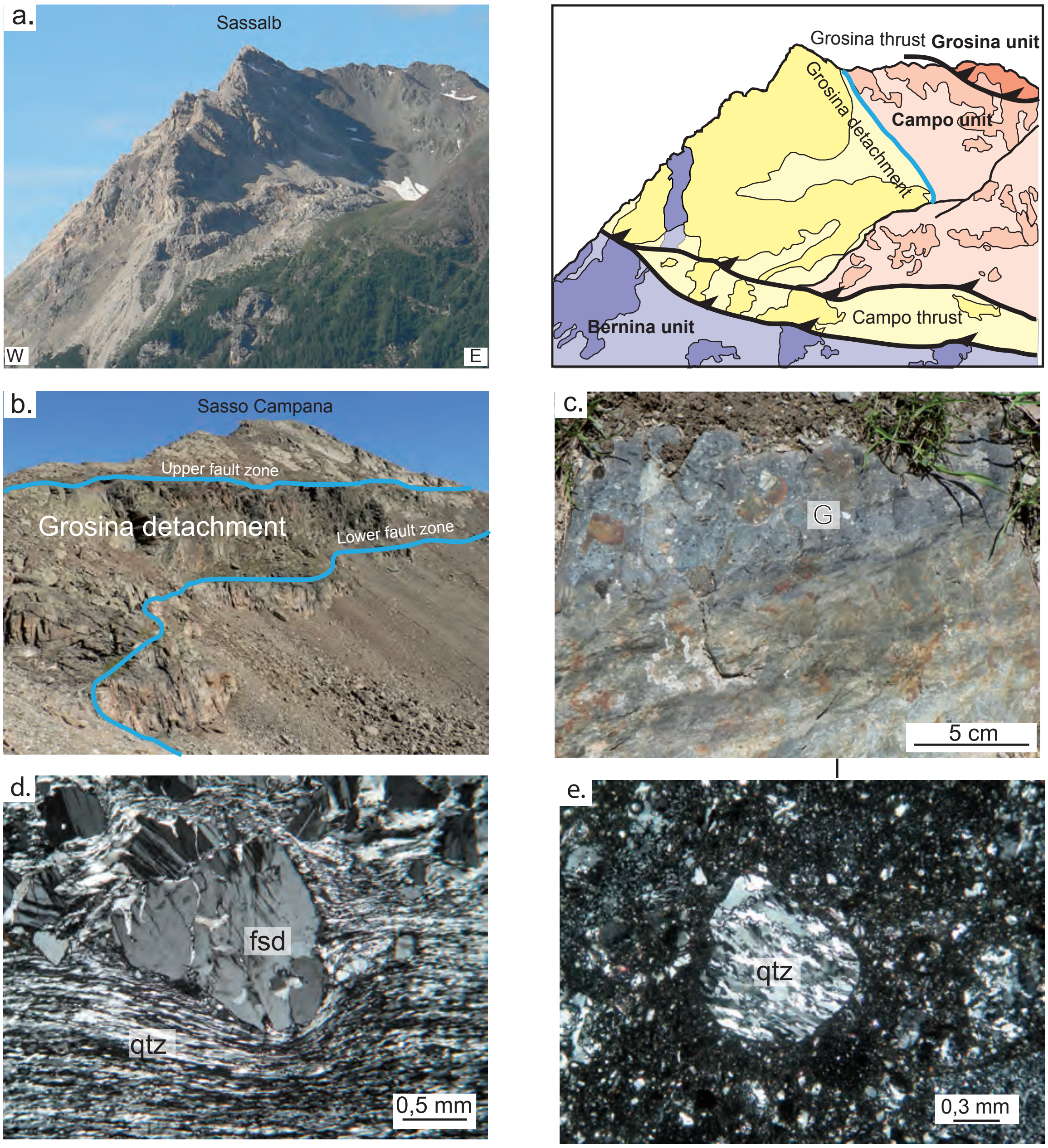


Fig. III. 9: Photographs and line drawing showing the Grosina detachment and related fault rocks (for localization of the photographs see Figs. 4a, 5b) (a) Contact between Bernina-Campo-Grosina units east of Val Poschiavo at Sassalb. Note that Campo unit is juxtaposed against the Bernina unit along a pre-Alpine contact referred to as Grosina detachment (in blue) cross-cut by the Grosina and Campo thrust faults. Note that the photograph has been inverted to be compared with the map (Fig. 4) and cross-section (Fig 5b) (804796/134771, Swiss topographic coordinates). (b) Grosina detachment at the top of Grosina unit in Sasso Campana. Note that two splay of the detachment can be observed (811306/137391, Swiss topographic coordinates). (c) Gouge zone as exposed along the Grosina detachment (811316/137314, Swiss topographic coordinates). Photographs of thin sections under crossed nicols. (d) Quartz mylonite from a shear zone that is cut by the Grosina detachment. Note the dynamically recrystallized quartz flows around feldspar clast (811311/137194, Swiss topographic coordinates). (e) Clast of quartz mylonite within a gouge from the Grosina detachment (811316/137314, Swiss topographic coordinates).

6. 2. Eita shear zone

The Eita shear zone is defined as the contact between upper crustal (Grosina) and mid crustal rocks (Campo). This structure can be mapped continuously over more than 20 km (Fig. 4) staying at the same structural level separating the Campo from the Grosina unit. This contact is well exposed in the Valle Grosina close to the village of Eita (Figs. 4a and 5b). North of Eita, within ten meters from the contact with the Grosina unit, the steeply-dipping Campo fabric is progressively overprinted by later deformation, leading to recumbent folds with E-W trending fold axes and a shallow W-dipping axial plane. A finite strain gradient is observed approaching the contact. This is best shown by the amplitude of recumbent folds associated with the rotation of the Campo fabric due to shearing along the Eita shear zone. Locally, Permian pegmatites are folded in the vicinity of the shear zone, indicating that shearing must post-date their intrusion (Fig. 10a). Microstructural observations reveal that the amphibolite-facies Campo fabric is dynamically retrogressed to greenschist facies conditions, as indicated by the break down of biotite to chlorite and destabilization of garnet and sillimanite into chlorite and white mica. The shear fabric is defined by quartz+white mica II+titanite associated with porphyroclastic K-feldspar, plagioclase and white mica I. Dynamically recrystallized quartz crystals show a strong lattice preferred orientation. Relics of porphyroclasts of quartz are displaying a core and mantle structures associated with small bulge along older grains of quartz typical for bulging recrystallization. Newly recrystallized fine quartz grains are locally arranged oblique in respect with the main foliation indicating also subgrain rotation recrystallisation. K-feldspar and plagioclase, on the contrary, undergo brittle deformation (Fig. 10c). Newly crystallized chlorite associated with fine grained white micas II show a preferred orientation, while large (1 mm) fractured and kinked white mica I is also oriented parallel to the mylonitic fabric. Stretching lineations defined by elongated K-feldspar porpyroclasts dip gently to the NNW. Shear bands, mica fish, and asymmetric strain fringes around K-feldpars indicate a top-to-the NNW sense of shear (Figs. 10 b, c). The evidence of brittle deformation of feldspar, associated with the growth of chlorite, white mica and the dynamic recrystallisation of quartz enable to constrain that shearing had to occur in the range between 250 and 400°C. Further to the west, the presence of green-brown biotite into the shear fabric, together with microstructural evidence of grain boundary migration in quartz, has been taken as an indicator that shearing had to occur at temperatures around 400°C (Meier 2003).

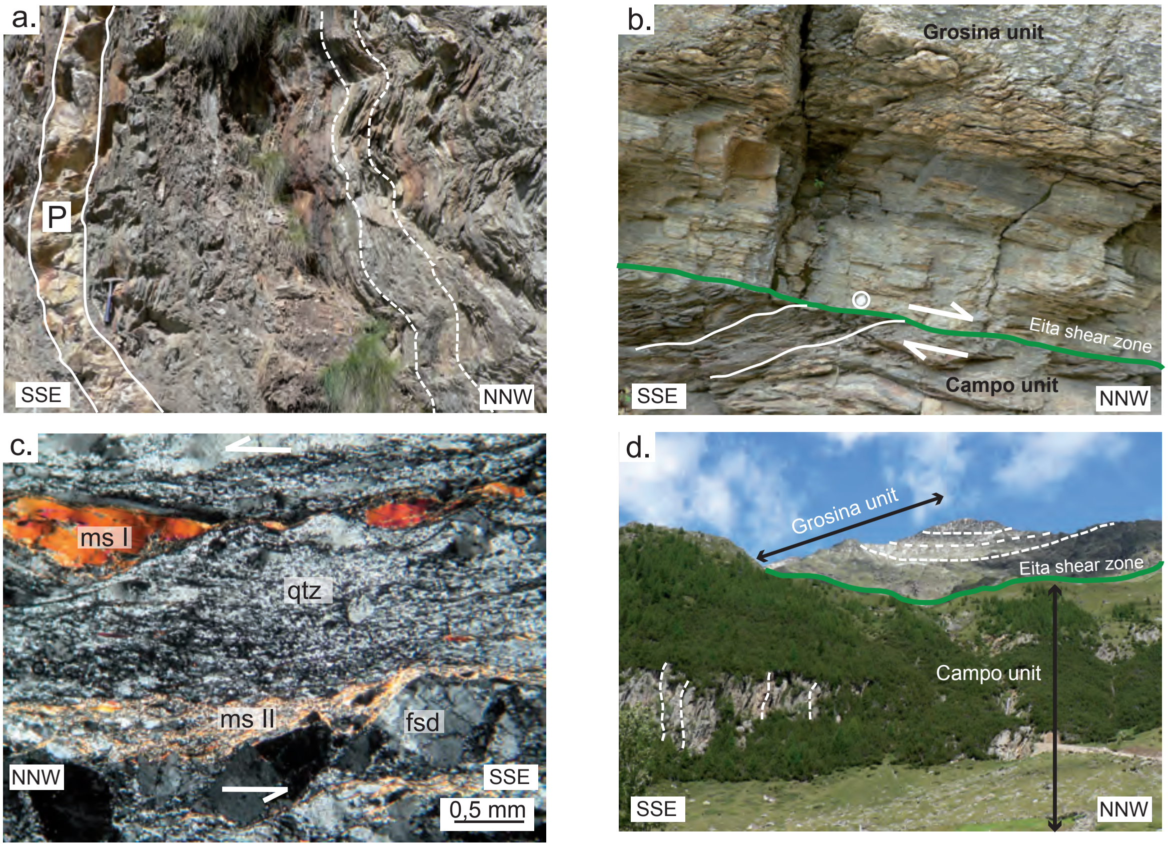


Fig. III. 10: Photographs of the Eita shear zone in Val Grosina near Eita (for localization of photograph see Figs. 4a and 5b) (a) Amphibolite-facies metapelites of the Campo unit cross-cut by Permian pegmatite (P). These two lithologies are subsequently folded in the vicinity of the Eita shear zone (816467/141107, Swiss topographic coordinates). (b) Eita shear zone, note the drag of the Campo fabric due to the shearing indicating a top the NNW sense of shear (816481/141150, Swiss topographic coordinates). (c) Photograph of a thin section under crossed nicols. Quartz is dynamically recrystallized while feldspar shows a brittle deformation. Sense of shear can be determined along C'-type shear bands indicating a top to the NNW sense of shear (816481/141150, Swiss topographic coordinates). (d) Large scale structures observed in Val Grosina: Note the angular discordance between the steep Pre-Permian Campo fabric and the flat pervasive greenschist pervasive foliation in the Grosina unit linked to the Eita shear zone (in green) which forms the contact between the Campo unit (footwall) and Grosina unit (hanging wall) (816305/139771; Swiss topographic coordinates).

In the first two hundred meters in the hanging wall of the Eita shear zone in the Grosina unit, older high-T fabrics and an intrusive granodiorit are overprinted by a foliation that is sub-parallel to the Eita shear zone (Fig. 10b). This fabric is defined by quartz+white mica II+chlorite+illmenite associated with porypyroclastic white mica I, K-feldspar and palgioclase, for which greenschist facies conditions of formation can be postulated. In many places, it can be observed that an older fabric defined by green biotite and porpyroclastic white mica I is observed in the microlithons between the new foliation defined by chlorite and newly crystallized small white mica II minerals. Quartz shows evidence of dynamic recrystallization into fine grained aggregates. New small quartz crystals are crystallizing along lobate porphyroclast boundaries indicating the presence of dynamic recrystallization processes. These observations confirm that the first two hundred meters above the Eita shear zone are retrogressed to greenschist facies metamorphic conditions with temperatures around 300 to 400°C.

6. 2. 1. $^{40}Ar/^{39}Ar$ geochronology of the Eita shear zone

In previous investigations, the Eita shear zone was interpreted as a Late Cretaceous Alpine thrust (Meier 2003). However, metamorphic conditions as well as structural orientation of the Eita shear zone are inconsistent with an Alpine origin of this structure. Indeed Late Cretaceous Alpine deformation is restricted to few, well-defined and localized tectonic horizons showing a top to the west sense of shear. In order to constrain the evolution and timing of this shear zone, we have taken one sample for $^{40}Ar/^{39}Ar$ dating on white micas. Localization of sample and descriptions are given in table 3 (for details about the analytical techniques see chapter 5, step-heating data are listed in table 2).

Sample	Host rock	Petrography of sample
Sample 17 822045/138708 (Swiss topographic coordinates)	Quartz mylonite	● Sample is defined by quartz+white mica +biotite locally retrogressed into chlorite ● Microstructure: Quartz are highly strained and display undulose and patchy extinction, deformation lamellae. Newly crystallized fine grained quartz crystals are developing along sutured poryphroclast boundaries, suggesting bulging recrystallization process. Porphyroclastic white micas are deformed by fracturation and kinking. Micas are elongated parallel to the foliation and are locally associated with fine-grained aggregates. ● Microstructures of quartz are indicating temperature of deformation between approximately 300°C and 400°C (e.g. Stipp et al, 2002).

Table 3: Sample location (with Swiss grid coordinates) and descriptions of the mineral assemblage.

6. 2. 2. Results of $^{40}Ar/^{39}Ar$ step-heating of white mica and significance of the obtained age (Fig. 11)

Sample 17 yields a "plateau" age of 203.37±1.26 Ma. The beginning of the age spectrum, starting from step 2, shows a slightly discordant shape comprised between a minimum age of 197 Ma and a maximum age of 205 Ma. Although two generations of white mica are found in the sample, the rare fine grained aggregates of white mica have been selected out by hand picking. Therefore, the apparent age spectrum is interpreted as resulting from the degassing of the argon reservoir related to the large porphyroclast white mica crystals. The age obtained from this sample is interpreted as a cooling age of the Eita shear zone.

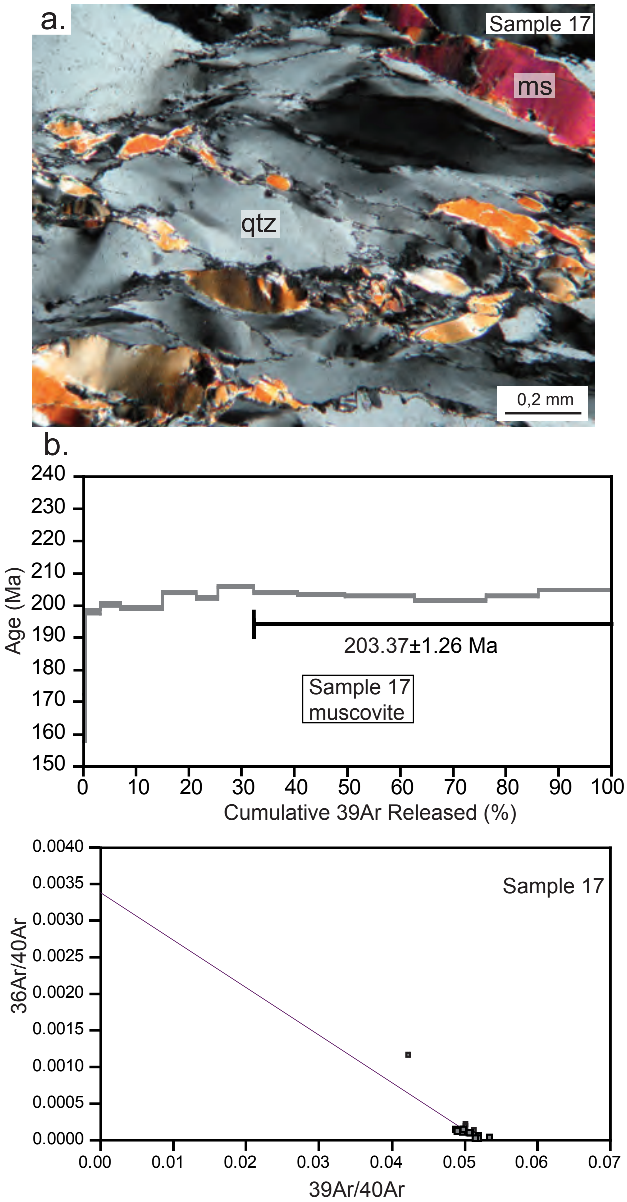


Fig. III. 11: Photomicrographs under crossed nicols of sample 17 that has been dated in the Eita shear zone. The sample corresponds to a quartz mylonite associated with deformed white mica (822045/138708; Swiss topographic coordinates) (For more details see table 2). Below are $^{40}Ar/^{39}Ar$ step-release sepectra and $^{39}Ar/^{40}Ar$ vs $^{36}Ar/^{40}Ar$ plot of the Sample 17 (for discussion of the results see text and for data see Table 2, 3).

To conclude, the Eita shear zone represents a low-angle shear zone which is located at the same structural level marking the limit between pre-rift upper crustal rocks (Grosina) and mid-crustal rocks (Campo). Several lines of evidence indicate that the Eita shear zone was active during Jurassic rifting: (1) it marks the limit between pre-rift upper (Grosina unit) and middle crust (Campo unit); (2) a new $^{40}Ar/^{39}Ar$ age on muscovite from this shear zone is coherent with the Jurassic cooling age of the Campo basement indicating exhumation of pre-rift middle crust during rifting; (3) it shows a retrograde equilibration of the hanging wall and footwall rocks at greenshist facies; and (4) the shear sense is inconsistent with that observed along Alpine thrust faults (top-to-the NW). Therefore we postulate that this structure was active during Jurassic rifting. However, in contrast to detachment faults that cut across brittle layers and result in the exhumation of the footwall relative to the hanging wall, the Eita shear zone stays in the same crustal level. Therefore, we interpret the Eita shear zone as a mid-crustal decollement that decoupled deformation in the upper crust from deformation in the lower crust. It is also important to note that the sense of shear identified along the Eita shear zone is not perpendicular to the former margin and the other Jurassic structures described in this paper. This may either be do to Alpine folding (in this case, the sense of shear would not represent anymore the original pre-Alpine sense of shear) or due to flow lateral to the margin in the middle crust. Further studies will be necessary to better understand the kinematic and tectonic significance of the Eita shear zone.

6. 3. Bernina detachment

The Bernina detachment represents a major structure that can be mapped over more than 20 km between Paleozoic basement rocks in the footwall and basement rocks, Permian rhyolites, pre-rift sediments or syn-rift sediments in the hanging wall. In most places, this fault is reactivated by subsequent Alpine thrust faults. The pre-Alpine structure is best preserved in the Val dal Fain area, where the structures in the basement and their relation to the overlying pre- and syn-rift sediments are preserved. In this area the detachment is characterized by a damage zone, 1 to 20 m thick that is formed by cataclasites. Towards the top of the basement a transition from cataclasites to foliated cataclasites can be observed. Mylonites are never observed associated with this detachment. Locally, matrix supported fault gouges with rounded,footwall-derived clasts can be observed (Fig. 12b). They are best developed where the top of the basement is formed by Permian rhyolites. Stretching lineations indicate an SW-NE direction and based on shear band and asymmetric clasts of K-feldspar a top to the south-west sense of shear can be determined (Fig. 12b).

The top of the fault is formed by a smooth surface with locale fissures filled by indurated tectono-sedimentary breccias (Figure 12a). The fault surface is either overlain by extensional allochthons consisting of upper crust and pre-rift sediments (Figs. 5c, 7) or, as in Val dal Fain, onlapped at a low angle (< 20°) by the syn-rift Allgäu formation. The fragments of pre-rift lying directly on the basement are made essentially by Upper Triassic sediments. Deeper parts of the pre-rift succession occur in bigger blocks, such as the Piz Alv allochthon (Fig. 8a). These slices of pre-rift are cross-cut by small scale high-angle normal faults, which are sealed by the overlying sediments. This complex geometry implies that the pre-rift succession was emplaced tectonically onto the basement and do not represent olistolites. This clearly indicates that the surface formed as a tectonic surface that was locally exposed at the seafloor. Therefore we call this structure a low angle or top basement detachment fault; e.g. Hoelker et al. 2003). The discontinued pre-rift slices as well as the exhumed basement are sealed by the Allgäu formation. It is important to note that where the detachment surface is onlapped by the Allgäu formation, neither gouges are observed, nor do basement clasts occur in the overlying sediments. This suggests that exhumation was followed by erosion and removal of the gouges and cataclastic basement, before the basement was onlapped by the Allgäu formation. Although in the direct neighborhood basement-derived breccias are not observed, they occur in the Middle Jurassic Saluver Formation in the more distal Err domain (Fig. 3).

In the Val dal Fain area, high-angle faults with vertical displacements ranging from 10 to 600 m cut the Bernina detachment (Figs. 5a and c). Because the basal sediments show a thickening towards these faults and the faults are sealed by the same sediments, it can be assumed that these high-angle faults formed during the sedimentation of the Allgäu formation. Thus, the sediments of the Allgäu formation were deposited after exhumation of basement along the Bernina detachment fault, but before the end of tectonic activity in this domain. These relationships confirm the Jurassic age of the Bernina detachment, which we interpret as a structure similar to the Err detachment exposed in the Err nappe (e.g. Froitzheim and Eberli 1990; Manatschal and Nievergelt 1997). Based on these new observations we propose a restored section across the Piz Alv-Val dal Fain area in which the Bernina detachment represents a low-angle detachment fault overlain by extensional allochthons that are sealed by syn- and post-rift sediments (Fig. 5c).

6. 4. Margna shear zone

Although the Margna shear zone has been intensively overprinted by the Alpine deformation, Herrmann and Müntener (1996) and Bissig and Hermann (1999) were able to demonstrate that this structure was active during Jurassic rifting and was responsible for the juxtaposition of uppermost crust against lowermost crust and subcontinental mantle (Fig. 7). Based on the work of these authors, a top to the southeast sense of shear has been proposed for the Margna shear zone. However, due to the strong Alpine overprint it is difficult to determine the exact metamorphic conditions and timing of this shear zone.

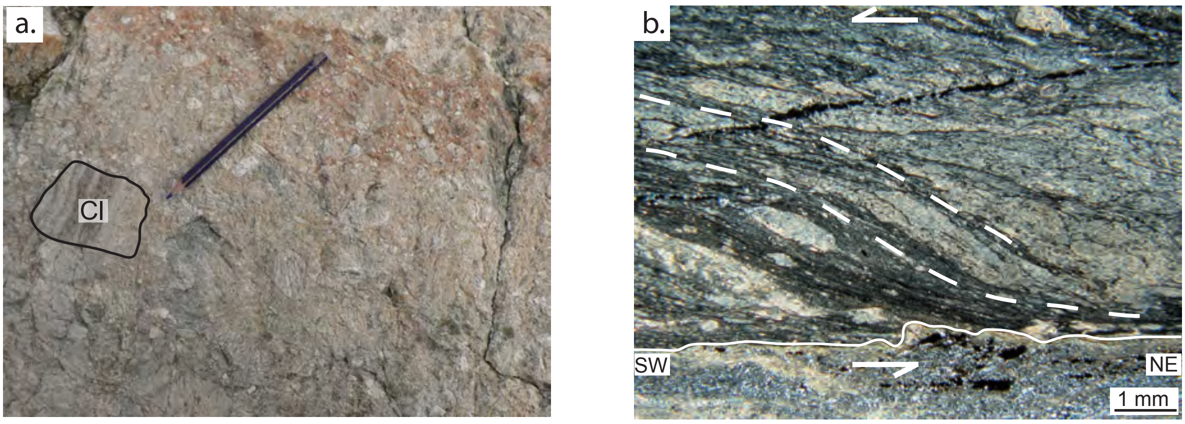


Fig. III. 12: (a) Photographs of the Bernina detachment, cataclastically deformed Paleozoic basement of the Bernina unit (794832/149133). (b) Photomicrographs under crossed polarizer of a gouge layer showing rotation of the shear fabric during shearing indicating a top of SW sense of shear (794643/143200) (Swiss topographic coordinates).

7. Discussion

Previous studies showed that the structures of the proximal margin are very different from those observed in more distal parts of the former rifted margin. This leads to the question about the nature of the crust within the necking zone separating the proximal and distal margin as well as the structures responsible for crustal thinning. In this study we focused on the Bernina-Campo-Grosina units, which lie in an Alpine tectonic position between units from the proximal and distal Adriatic fossil margin. In order to define the necking zone, we first need to characterize the nature and structure of the crust within the proximal and distal margins (Fig. 13).

Relics of the proximal margin are best exposed within the Upper Austroalpine Ortler unit as well as in the Lombardian basins in the Southern Alps (Fig. 3). These units are mainly formed by a Permo-Mesozoic cover overlying pre-rift upper crust. Rift structures as well as the associated stratigraphic evolution are very similar from those observed at present-day proximal margins (e.g. Jeanne d'Arc basin; Newfoundland margin). Using refraction and reflexion seismic data it could be demonstrated in present-day margins that these types of basins are not related to major thinning of the crust. Based on these analogies, we assume that the relics of proximal units in the Alps derive from a relatively little thinned and undisturbed crustal section. In contrast units from the former distal margin exposed in the Err and Margna units in the Eastern Alps and the Canavese and Ivrea-Verbano- Strona-Ceneri units in the Southern Alps show a very different structural and stratigraphic evolution. These domains are made exclusively of relics of felsic pre-rift upper and lower crust intruded by mafic rocks during the Permian, while mid-crustal levels are missing. Because these units preserve direct relationships with units consisting of exhumed subcontinental mantle (Platta and Malenco units) and because they are overlain by deep water sediments (e.g. Radiolarian cherts and Calpionella limestone), they were interpreted, in analogy with the Iberia rifted margin as hyper extended distal margins (Manatschal 2004). In such hyper extended rifted margins, the crust is typically less than 10 km thick and thins oceanwards to zero, as indicated by the occurrence of mantle rocks exhumed at the seafloor covered by post-rift sediments.

In contrast to the proximal and distal rifted margins, for which the nature and structure of the crust are reasonably well described by drill hole as well as refraction and reflection seismic investigations, only little is known about the necking zone at the transition between the proximal and distal margins. Based on detailed mapping combined with a structural interpretation and dating of the Bernina-Campo-Grosina units, which are lying between proximal and distal units of the former margin, we are now able to identify primary pre-Alpine structures of the necking zone and to restore the first order structure of the necking zone.

A major pre-Alpine structure, the Grosina detachment, juxtaposes the units from the distal margin (Err/Bernina/Margna) against the necking zone (Campo-Grosina) (Figs. 5c and 13). The Campo-Grosina domain is at present formed by a 3 km thick crustal section consisting of pre-rift upper and mid crustal rocks. The Err-Bernina-Margna domain represents remnants of the former distal margin consisting of pre-rift upper and lower crustal rocks but not mid-crustal rocks (Fig. 13). These observations suggest that the necking zone is representing a major boundary separating two distinct parts of the margin characterized by a different crustal structure and stratigraphic evolution as indicted by fromer studies of the proximal and distal margins.

7. 1. Deformation structures associated with extreme crustal thinning

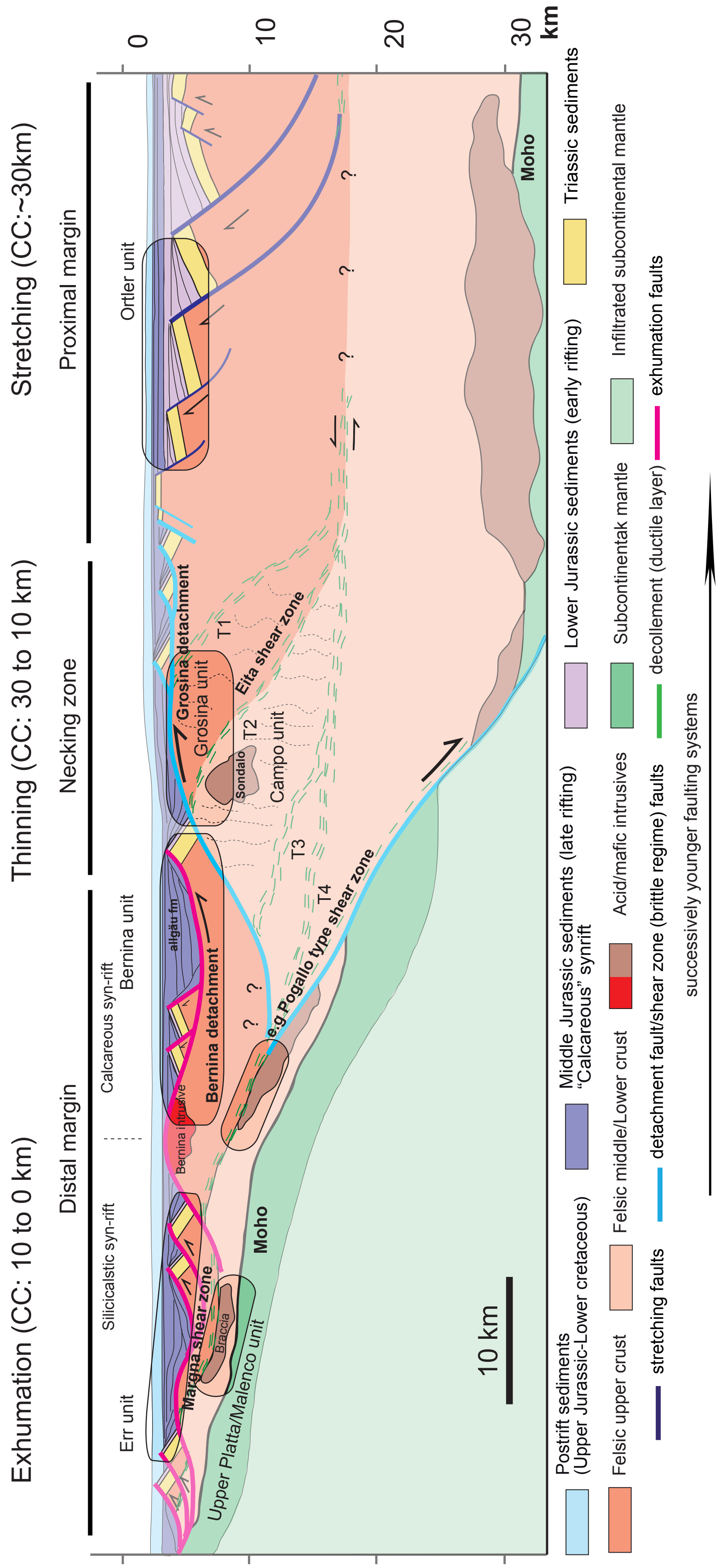
In the Campo-Grosina necking zone and in the Bernina-Margna distal margin, several diachronous extensional structures can be mapped (Fig. 4a). These structures can explain crustal attenuation and exhumation of the mid-crust in the necking zone associated with juxtaposition of upper against lower crust in the distal margin and final mantle exhumation in the OCT. Similar structures were also described from the Canavese and Ivrea-Verbano- Strona-Ceneri units in the Southern Alps. Because the Eastern and Southern Alps sections are preserving different crustal portions of the fossil Adridatic rifted margin and were reactivated in a different way, these two sections provide complementary information on how extreme crustal thinning was accommodated during Jurassic rifting. A major result of our study is that crustal thinning results from the interplay between several decoupling shear zones and brittle faults that strongly depend on the rheological evolution of the extending lithosphere. Below we describe major structures that can be identified in the necking zone of the Adriatic rifted margin.

7. 1. 1. The Eita shear zone: a mid-crustal decollement

The Eita shear zone represents a crustal shear zone separating pre-rift middle crust (Campo unit) from pre-rift upper crust (Grosina unit). This structure marks a continuous ductile shear zone, which can be followed in the field over more than 20 km (Figs. 4a and 5b). In the footwall of the Eita shear zone Permian and/or pre-Permian high-T fabrics are retrogressed and overprinted by a pervasive greenschist facies deformation which is also overprinting the first 200 m of the overlying Grosina unit. The mylonitic retrograde greenschist facies fabric is defined by an intense dynamic recrysatillsation in quartz associated with bulging and subgrain rotation recrystallization but still brittle behavior of feldspar. Based on metamorphic condition and the observed fabrics it can be proposed that the Eita shear zone was active between 300 and 400°C. These conditions indicate that this structure was probably active at 10 to 15 km depth at the brittle ductile transition. Since this structure can be mapped over 20 km within the same crustal level showing the same metamorphic conditions and quartz fabrics, we assume that this shear zone may have been active as a decollement decoupling deformation between rigid upper and mafic lower crustal and upper mantle layers .

Stretching lineations along the Eita shear zone are indicating a NNW sense of shear, which is oblique in respect with NW-SE Alpine thrust direction as well as E-W directed rift directions. This observation indicates an oblique component for the extension of the shear zone in respect with other rift structures. How this sense of shear can be interpreted in the overall kinematic framework of the rift evolution is yet unclear.

The age of the Eita shear zone is constrained by $^{40}Ar/^{39}Ar$ cooling ages in the Campo unit as well as by dating of mylonitic fabrics which gives a late Triassic to early Jurassic age (~203 Ma). Several similar shear zones have been identified in the Campo and Grosina units that were however, not yet dated. Although their exact role and age are not yet determined we do not exclude that these shear zones may belong, together with the Eita shear zone to a network of shear zones that were distributed in the ductile middle crust.



Exhumation (CC: 10 to 0 km)
Thinning (CC: 30 to 10 km)
Stretching (CC:~30km)
Distal margin
Necking zone
Proximal margin
Err unit
Silicicalstic syn-rift
Calcareous syn-rift
Bernina unit
Ortler unit
Margna shear zone
Braccia
Upper Platta/Malenco unit
Moho
Bernina intrusive
Bernina detachment
allgäu fm
e.g Pogallo type shear zone
T1
T2
T3
T4
Grosina detachment
Grosina unit
Eita shear zone
Sondalo
Campo unit
10 km
0
10
20
30
km
Postrift sediments (Upper Jurassic-Lower cretaceous)
Middle Jurassic sediments (late rifting) "Calcareous" synrift
Lower Jurassic sediments (early rifting)
Triassic sediments
Felsic upper crust
Felsic middle/Lower crust
Acid/mafic intrusives
Subcontinentak mantle
Infiltrated subcontinental mantle
stretching faults
detachment fault/shear zone (brittle regime) faults
decollement (ductile layer)
exhumation faults
successively younger faulting systems

7. 1. 2. Grosina detachment: a major thinning structure

The Grosina detachment is best exposed at the top of the Grosina unit (Figs. 4, 5b and 9b). This detachment is an intra basement low-angle detachment fault, which can be mapped at the top of the Grosina unit over 8 km and can be correlated westwards separating the Grosina-Campo units in the footwall from the Bernina unit in the hanging wall (Fig. 5b). At Sassalb (Fig. 9a) exhumed mid-crustal rocks (Campo unit) are juxtaposed along the Grosina detachment against pre-rift Triassic dolomites and upper crustal rocks of the Bernina domain. In contrast to the Eita shear zone, the Grosina detachment was an exhumation fault along which rocks from deeper custal levels were exhumed and juxtaposed against upper crustal rocks and Mesozoic sediments. Micorstructural observations show that the Grosina detachment fault was active in the brittle field below ~300°C as indicated by the lack of evidence for dynamic recrystallization and crystal plastic deformation in quartz. This structure is interpreted to be active during Jurassic rifting based on cross-cutting relationships with Permian structures and Alpine thrust faults. However within this brittle detachment, clasts of higher metamorphic grade can be found resulting from the reworking of mylonitic rocks. The cross-cutting relationship between the brittle Grosina detachment and this quartz mylonites is occurring with a discordance of about 20°. Three main possibilities can be inferred: 1) the quartz mylonites formed in a deeper part of the Grosina detachment and were exhumed and overprinted by brittle deformation, or 2) the quartz mylonites belong to a structure similar to the Eita shear zone, i.e. they represent a decollement that was exhumed along the detachment fault (e.g. mylonitic front in the Wipple mountains Lister and Davis 1989), or 3) the quartz mylonites are unrelated to Jurassic rifting and are older Permian or Variscan structures. At the moment the exact significance of this quartz mylonite remains unclear.

7. 1. 3. "Pogallo"-type shear zones: major ductile thinning structures

Unfortunately, more proximal and deeper parts of the necking zone are not exposed in the Campo-Grosina units. As discussed above, in the most distal part of the Southern Alps section, in the Canavese and Ivrea-Verbano - Strona-Ceneri units, similar structures are observed juxtaposing pre-rift upper and lower crustal levels (Fig. 3; section C-C') (for references see also Handy et al, 1999). Of particular interest is the Pogallo shear zone (Pogallo ductile fault zone of Handy 1987). This structure corresponds to an extensional structure responsible for the

Fig. III. 13: Tentative palinspastic reconstruction of the Adriatic rifted margin across the Austroalpine and South-Penninic units in Grisons (SE Switzerland). The "Pogallo type" shear zone as been added to the section in analogy to observations reported from the Southern Alps section C-C' (Fig. 3c). The frames shown in the figure correspond to observed relationships.

exhumation of the pre-rift lower crust exposed in the Ivrea-Verbano zone against pre-rift upper crust exposed in the Strona-Cenri zone. The Pogallo shear zone represents a km-wide moderate to low-angle shear zone, which can be followed over several km showing the variation of metamorphic condition of this structure from mid-amphibolite (650°C) to upper greenschist facies conditions (450°C) (Handy 1987). In previous studies this structure has been interpreted to be responsible for the exhumation of the Ivrea-Verbano lower crustal rocks to a depth of 10 km. Based on $^{40}Ar/^{39}Ar$ data on white mica and biotite Handy (1987 and references therein) suggested that this fault was active between 210-180 Ma. More recent work by Mulch et al. (2002) gave an age of 173±4 Ma resulting from a discontinuous retrograde metamorphism associated with crustal thinning along the Pogallo ductile shear zone. Thus, in contrast to the Eita shear zone, the Pogallo shear zone is exhuming and thinning the crust, and in contrast to the Grosina detachment, it is active in the lower crust. However, all these structures are active during early Jurassic rifting and are within a position between the proximal and distal Adriatic rifted margin. Because the Grosina fault only affects brittle layers (lack of mylonites along this fault) and the Pogallo shear zone does not daylight (see Manatschal et al. 2007), we assume that these faults are soling out at mid-crustal levels within decoupling horizons. Thus, based on our observations we interpreted that crustal thinning is accommodated by a system of conjugate crustal detachment structures that are active in the upper brittle crust (e.g. Grosina detachement fault) and in the mafic lower crust and upper mantle (e.g. Pogallo type shear zone) and are decoupled along mid crustal decollements (e.g. Eita shear zone) (Fig. 13).

7. 1. 4. Margna shear zone

Along the Margna shear zone, first described by Hermann and Müntener (1996), pre-rift lower crustal rocks preserving a primary contact with subcontinental mantle are juxtaposted against pre-rift upper crust preserving a primary contact to Mesozoic sediments. Hermann and Müntener (1996) and Bissig and Hermann (1999) interpreted this structure as a Jurassic contact based on structural observations. Because of the strong Alpine tectonic overprint, it is difficult to restore the geometry and the metamorphic conditions of this shear zone as well as to link it with previous described structures. However, this shear zone is located at the interface between uppermost and lowermost crust with more than 20 km of continental crust missing inbetween. As discussed below and shown in Fig. 14, we suggest that this structure represents a complex shear zone that forms an extraction shear zone in front of the necking zone (e.g. extraction fault of Froitzheim et al. 2006)

7. 1. 5. The Bernina detachment: an exhumation fault

The Bernina detachment is interpreted as a low-angle detachment fault accommodating the final extension in the distal margin. It can be mapped over more than 20 km across the entire Bernina unit at the top of the basement between upper crustal rocks and Permian rhyolites and/or pre- or syn-rift sediments. In the Val dal Fain and Piz Alv area (Fig. 8), the first phase of rifting is characterized by high-angle normal faulting associated with the deposition of the Alv breccias. The high-angle normal faults of this earlier rift phase have been subsequently cross-cut by younger low-angle detachment faults. This results in the formation of extensional allochthons that range in scale from several tens of kilometer wide blocks, including basement, to small, extended fragments of pre-rift dolomites overlying exhumed basement (e.g. Val dal Fain; Fig. 8b). Along the Bernina detachment no variation of the metamorphic conditions can be observed and the structures related to the detachment are forming in the brittle field at temperature below 300°C over more than 20 km along strike. This observation suggests a low-angle geometry of this fault, forming at least locally an exhumed, top basement detachment fault. This is confirmed by the low-angle between the fault zone and the syn-rift sediments onlapping onto the fault (Fig. 8b). The age of this structure is constrained by the deposition of the syn-rift Allgäu formation, which is Late Sinemurian to Bathonian in age indicating that this detachment was not younger than Bathonian. Similar structures in the adjacent Err and Platta units have been recognized and described by Froitzheim and Eberli (1990); Manatschal and Nievergelt (1997). In these areas, a similar detachment fault could be mapped as a coherent unit over 17 km, and restored over more than 30 km. The Bernina detachment is interpreted as a similar structure like the Err detachment, however, within a more proximal part of the distal margin. These structures accommodated the extension in the most distal continental margin, similar to the structures described from the present day Iberia distal margin, which are responsible for mantle exhumation.

7. 2. The temporal and spatial evolution of crustal thinning: an Alpine perspective

To unravel the temporal and spatial evolution of crustal thinning along the fossil Adriatic rifted margin, two major data sets need to be taken in account: the stratigraphic record and the ages obtained from the basement units and the structures accommodating crustal extensions. While the startigraphic record can be deduced from the whole margin system, including the conjugate European/Briançonnais margins, (for a review see Mohn et al. 2010), the exhumation and thinning in the basement can only be determined from locations where the Alpine tectonic and metamorphic overprint was not too high and rift structures are still preserved. Based on the correlation between stratigraphic, petrological and thermochronological studies the following spatial and temporal evolution of rifting can be proposed for the Adriatic margin.

7. 2. 1. Initial condition at the onset of rifting and importance of inheritance (Figs. 14a)

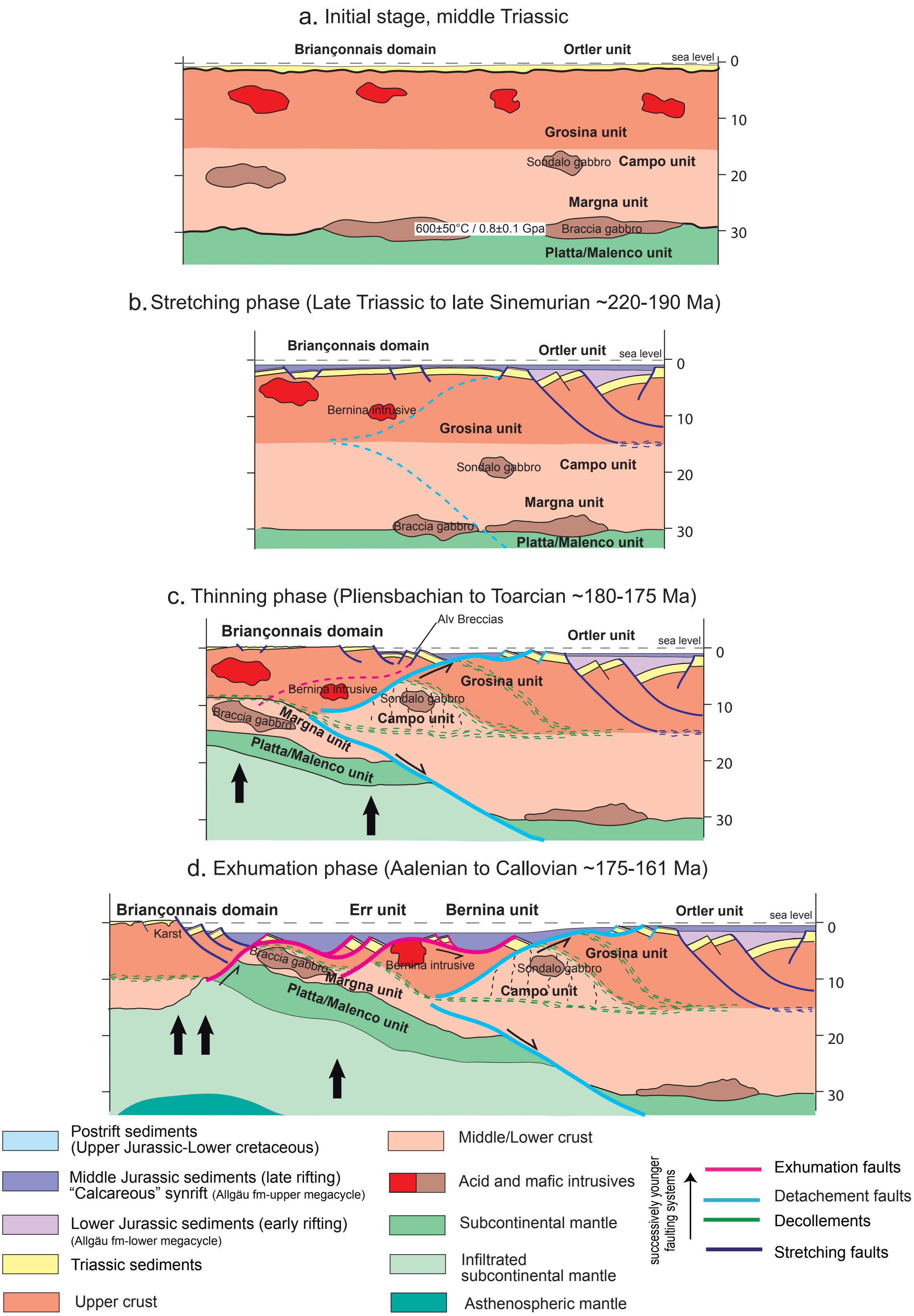
Pre-rift setting of the future Adriatic margin and Briançonnais domain during the Triassic has been characterized by the distribution of shallow marine carbonate platforms which suggest that the pre-rift crust was thermally and isostatically equilibrated. This is confirmed by petrological work from the crust-mantle section preserved in the Margna units indicating that before rifting the crust was equilibrated at 30 km as indicated by 600±50°C at 0,8 Gpa, suggesting a average heat flow of 70W/m2 (Müntener et al, 2000).

7. 2. 2. From stretching to thinning: birth of the necking zone (Figs. 14b and c)

First extensive structures related to Jurassic rifting are identified to be active at late Triassic to Late Sinemurian time (~205-190 Ma). These structures form large basins bounded by high-angle normal faults (e.g. Ortler unit; Eberli 1988). In the Southern Alps, Bertotti (1991) identified the listric geometry of Jurassic normal faults and demonstrated that these faults sole out at greenschist facies conditions between 10-15 km at mid crustal levels. This suggests that during early rifting deformation in the upper crust was decoupled from deformation in the mafic lower crust and subcontinental mantle. Moreover, exhumation of rocks to the seafloor, was, apart from fault scarps, a fortiori not possible. This, together with the overall subsidence of the margins explains that no basement is yet exhumed and reworked in the stratigraphic succession of the Alpine Tethys margins during this early stage of rifting (Fig. 14b). The same style of deformation is also observed on the European margin (e.g. Bourg d'Oisans, Lemoine et al. 1986). Attempts to quantifying the extension accommodated in the upper crust results in β-factors ranging between 1,2 and 1,65 (Froitzheim 1988; Bertotti et al. 1993). The conjugate European domain shows similar values for the same period (Chevallier et al. 2003).

From Pliensbachian to Toarcian (190-175 Ma) extension starts to shift towards the future distal margin (Froitzheim and Eberli 1990) resulting in a major change in the structural and stratigraphic evolution of the whole margin (Fig. 14c). In the Bernina domain, this period is characterized by the depostion of the Alv Breccias in fault bounded basins linked with the deposition of the second megacycle of the Allgäu formation. During Toarcian to Aalenian time (180-170 Ma) the Briançonnais domain, i.e. the future conjugate margin, undergoes a different and decoupled evolution from the rest of the margin as indicated by an uplift leading to a subaerial exposure, which contrasts from the subsidence recorded on the adjacent proximal margins. The subarial conditions will prevail until the Bathonian-Oxfordian (165-155 Ma), when the Briançonnais domain will subside rapidly.

This distinctive isostatic evolution is important to take into account, since it occurred at the same time as crustal thinning documented in the Ivrea-Verbano zone and Margna-Grosina-Campo domains. New ^{40}Ar/^{39}Ar data on biotite from the Campo basement, sampled in the vicinity of a Permian intrusion (e.g. Sondalo Gabbro) demonstrate that these rocks were exhumed and cooled below 320-350°C at around 180-185 Ma. Similar data have been described from the Ivrea-Verbano and the Margna units both intruded by Permian gabbros in a lower crustal position (~ 30km) (Handy et al, 1999 and references therein; Mulch et al, 2002; Müntener et al, 2000; Villa et al, 2000). P-T-t paths from the Ivrea-Verbano zone have shown an exhumation of pre-rift lower crust below temperatures of 350-320°C at 210-180. Therefore exhumations of deeper levels of the continental crust were contemporaneous with uplift of the Briançonnais domain, as indicated by the occurrence of karst and subarial sediments. The evolution of this domain during major crustal thinning is at odd with the classical ideas of McKenzie (1978), but however, commonly observed in hyper extended rifted margins (e.g. Iberia margin; Péron-Pinvidic and Manatschal 2009; South Atlantic and NWAustralia margin; Driscoll and Karner 1998). Because crustal attenuation was not directly accompanied by subsidence but linked to uplift (e.g. Briançonnais domain), we think that crustal thinning had to occur simultaneous with lithospheric thinning. Lithospheric thinning is documented in the Alpine system by the impregnation of magma into subcontinental mantle prior to its exhumation at the seafloor (e.g. Piccardo 2008; Müntener et al. 2009 and references therein). The magma-infiltration may have enhanced the thermal erosion, resulting in depth depended thinning of the lithosphere (e.g. Karner and Kusznir 2007 and Cannat et al. 2009).


a. Initial stage, middle Triassic
Briançonnais domain
Ortler unit
sea level
Grosina unit
Sondalo gabbro
Campo unit
Margna unit
600±50°C / 0.8±0.1 Gpa
Braccia gabbro
Platta/Malenco unit
0
10
20
30
b. Stretching phase (Late Triassic to late Sinemurian ~220-190 Ma)
Briançonnais domain
Ortler unit
sea level
Bernina intrusive
Grosina unit
Sondalo gabbro
Campo unit
Margna unit
Braccia gabbro
Platta/Malenco unit
c. Thinning phase (Pliensbachian to Toarcian ~180-175 Ma)
Briançonnais domain
Alv Breccias
Ortler unit
sea level
Bernina intrusive
Grosina unit
Sondalo gabbro
Campo unit
Braccia gabbro
Margna unit
Platta/Malenco unit
d. Exhumation phase (Aalenian to Callovian ~175-161 Ma)
Briançonnais domain
Err unit
Bernina unit
Ortler unit
sea level
Karst
Braccia gabbro
Bernina intrusive
Grosina unit
Sondalo gabbro
Campo unit
Margna unit
Platta/Malenco unit
Postrift sediments (Upper Jurassic-Lower cretaceous)
Middle Jurassic sediments (late rifting) "Calcareous" synrift (Allgäu fm-upper megacycle)
Lower Jurassic sediments (early rifting) (Allgäu fm-lower megacycle)
Triassic sediments
Upper crust
Middle/Lower crust
Acid and mafic intrusives
Subcontinental mantle
Infiltrated subcontinental mantle
Asthenospheric mantle
successively younger faulting systems
Exhumation faults
Detachement faults
Decollements
Stretching faults

7. 2. 3. From thinning to exhumation (Fig. 14d)

The thinning stage of the Alpine Tethys rift system, which occurs around Pliensbachian to Toarcian time, i.e. 190 to 175 Ma, is well documented in the stratigraphic record by a wide spread karst and a major unconformity in the Briançonnais domain and in the basement by exhumation of deeper crustal level as documented by numerous cooling ages (e.g. Ivrea, Campo and Marnga units). These events pre-date the exhumation of mantle rocks to the seafloor dated by radiolarian cherts and U/Pb on zircons from oceanic gabbros at 165 to 160 Ma (Bill et al. 1997; Schaltegger et al. 2002). Final rifting is characterized by the localization of deformation in the future distal margin over previously thinned crust (Fig. 14d). In contrast to the previous stage, the new faults cut through a thin, brittle crust and are able to penetrate into the underlying subcontinental mantle. These new structures form top basement detachment faults that result in the delamination of the hanging wall and the exhumation of upper crust and eventually also lower crust leading finally of exhumation of subcontinental mantle at the seafloor. The top to the west (ocean) sense of shear determined from these detachments are in line with the fact that these detachment faults form break-aways to the east (continent) and cut oceanwards into mantle. The allochthons occurring over the detachment faults (e.g. Val dal Fain – Piz Alv area; Figs. 5a and 7), are interpreted to derive from the delamination of the hanging wall block, i.e. the Briançonnais domain. This interpretation is supported by the similarity between the pre-rift stratigraphic record of the Briançonnais and distal Adriatic margins.

The formation of exhumation surfaces is best recorded in the stratigraphic record with the occurrence of reworked crustal rocks in the syn-rift sediments, which led to the creation of a new silisiclastic sediment source. In the Southern Alps in the Canavese zone, syn-rift sediments even show the occurrence of clasts of lower crustal rocks of the Ivrea-Verbano zone (Ferrando et al. 2004).

Fig. III. 14: Conceptual model showing the evolution of lithospheric thinning as recorded in the Adriatic fossil margin: (a) the pre-rift situation in middle Triassic, (b) stretching phase from late Triassic to Sinemurian (220 to 190Ma), (c) thinning phase from Pliensbachian to Toarcian (190 to 175Ma), (d) exhumation phase from Toarcian to Callovian (175 to 161Ma). Note the spatial and temporal evolution of the Bernina intrusive, Sondalo and Braccia Gabbro. For more details and further discussion see text.

7. 3. Strain evolution and strain partitioning between brittle and ductile layers (Fig. 15)

The main question, which is also at the origin of this paper, is how the crust thins and what are the structures that can account for extreme crustal thinning. In Figure 14 and in the previous section, we describe the main geological observations and propose a geological model showing the temporal and spatial evolution of the margin, focusing to the evolution of the necking zone and the crustal thinning. The aim of figure 15 is to discuss the strain evolution and the structures that accommodate crustal thinning in a more conceptual way. In order to simplify figure 15, we assume a simple four layer rheology where inheritances as well as the initial stretching phase (formation of rift basins during an initial stage of rifting) were not considered. The concept shown in figure 15 is based on two key assumptions: 1) mass needs to be conserved during rifting (problem of restoration of sections), and 2) deformation in brittle and ductile layers needs to be partitioned (crust does not only behave as brittle or ductile layer). The aim is to discuss the first order processes of crustal thinning and mantle exhumation and the temporal and spatial relationships between the major structures discussed in this paper.

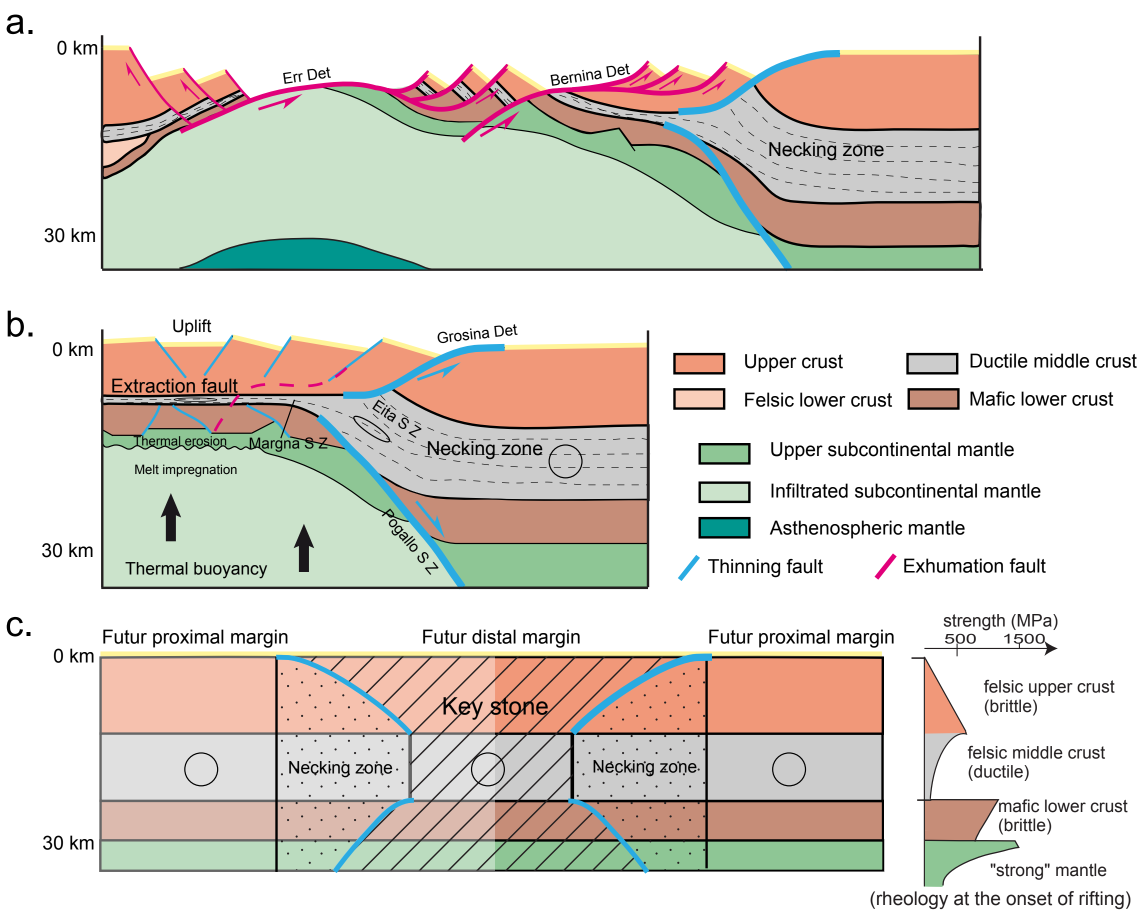


Fig. III. 15: Cartoon showing the strain distribution and strain partitioning during lithospheric thinning: (a) exhumation phase; (b) thinning phase; (c) initial stage. For a discussion see text.

As shown in Fig. 15 mantle exhumation is post-dating crustal attenuation and complete embrittlement of the crust, indicating that these processes are polyphase and the result of the interplay between structures active at all crustal layers.

Onset of thinning occurs when deformation starts to localize within the future distal margin. In Fig 15c, this is shown by a conjugate set of detachment faults that are in the upper felsic brittle crust and that are decoupled in a ductile layer in the ductile felsic middle crust. Extension in the rigid mafic lower crust and upper mantle is accommodated along conjugate shear zones. This set of conjugate detachment faults limits the future distal margin from the adjacent proximal margins, and define the necking zones of the margin.

During subsequent thinning, only the hatched areas (necking zones plus future distal margin) will deform. Brittle layers will deform by faulting and block rotation, while ductile layers will deform by homogeneous strain (see strain ellipsoid). A more complex process is supposed to occur in the mantle, since brittle fracture and boudinage of the brittle upper mantle may induce uplift and replacement by deeper, hotter impregnated mantle. In term, this thermal erosion could be at the origin of the uplift of the future distal margin (e.g. key stone, Briançonnais) during final rifting (for more discussion see Piccardo 2008; Müntener et al. 2010 and references therein). Crustal thinning is accommodated by conjugate detachments in the brittle layers shown in blue (e.g. Grosina and Pogallo) that are limited by a ductile layer which acts as a decoupling horizon between the two brittle layers. In the ductile layer strain will increase towards the necking zone (e.g. strain ellipsoids) resulting in shear zones (e.g. Eita shear zone). The ductile layer within the future distal margin will act as an extraction fault (channel), along which the ductile layer will be thinned in-between the rigid upper and lower crust/mantle (e.g Margna shear zone).

The transition from symmetric thinning to asymmetric exhumation occurs when ductile mid-crustal layers cannot decouple anymore the deformation in the upper crust and mafic lower crust and subcontinental mantle. At this stage faults can cut from the surface into mantle and result in the exhumation of mantle rocks in the OCT. These faults (shown in red), are typically downward concave faults that sole out at shallow levels in the mantle and are expressed as low-angle top basement detachment faults in the distal margin and adjacent OCT (e.g. Bernina and Err detachment faults). During this stage of deformation the former key stone (Briançonnais or H-block of Lavier and Manatschal 2006) will be delaminated and will form units constituting the final distal margin.

7. 4. Implications for present-day rifted margins

At present-day rifted margins necking zones are yet little explored. However thanks to the increasing number of high quality seismic reflection and refraction data, the nature of the basement and the first order structures of the necking zones can be imaged, however, their interpretation remains difficult do to a lack of scientific drill hole data from the basement of the necking zone. In figure 16, we show the refraction seismic line IAM5-Sines across the Iberia rifted margin from (Afilhado et al. 2008) to exemplify the first order crustal structure of the margin in order to compare it with the structure observed and discussed in this paper. In the IAM5 section (Fig. 16a), the necking zone corresponds to the domain where the curst is thins from ±30 km to 10km. The Moho geometry can be found below the necking zone. It is characterized by steeply dipping surfaces that can be up to 35° (Lau et al. 2006). Beneath the distal margin, the Moho is flattening and disappears near the OCT due to serpentinization within the mantle (Minshull 2009), while in the distal margin, strong reflections occur within the layer with velocities ≤ 8km/s, which is commonly interpreted as the contact between thinned crust and serpentinized mantle (e.g. "s" reflection of Reston 1996). Further continentwards, in the necking zone, strong sub-horizontal reflections occur, which are clearly within the crust. These reflections are commonly interpreted as "Conrad" reflections, i.e. the contact between upper and lower crust, which corresponds usually to a brittle ductile transition. In the section mid crustal velocities seem to disappear in the necking zone and in the most distal margin only upper crustal velocities are observed. Therefore, Afilhado et al. (2008) suggested that the middle crust is wedging out towards the necking zone. Indeed, our restoration from this margin (Fig. 13) is showing many similarities with the refraction seismic data shown in the IAM5 section (Fig. 16a). The first order structure of the crust, i.e. the wedging out of the middle crust, the juxtaposition of upper and lower crust as well as exhumation of mantle rocks are well recorded in the relics of the Adriatic margin in Alps. Therefore, we assume that the underlying processes and the strain accommodation within the two margins can be compared as well. This enable us to propose a coherent structural interpretation of some of the reflections observed in the section in Fig. 16b. The Margna-Pogallo shear zone could be linked to the contact between middle crust and mantle that dip continent wards. Many authors have proposed that the seismic Moho at the base of the necking zone may correspond to a shear zone (e.g. Péron-Pinvidic and Manatschal 2009). The apparent dip of the Pogallo shear zone was estimated by Handy (1987) between 10° to 34° which is coherent with observation in present day margins. The high velocity bodies that seem to be displaced by the Moho could in this case correspond to underplated gabbroic bodies similar to those observed in the Margna and Ivrea units.

The Eita shear zone and associated shear zones in the Campo unit could be equivalents of the reflections in the IAM5 section (Fig. 16a). These reflections are like the Eita shear zone at the top of middle crust and point oceanwards towards the wedge of the middle crust in the necking zone.

The Grosina and Bernina detachments are top basement detachment faults, locally overlain by extensional allochthones. These structures are therefore not easy to image and interpret in seismic sections (e.g. Hölker et al. 2003). We suggest, however, that an equivalent structure to the Grosina detachment may occur east of the B1 high (see Fig. 16a). This is supported by the observation that the middle crust wedges out in the footwall of this structure and that this structure also separates the distal and proximal margins showing different basin architectures.

The similarity between the structures observed in the Bernina, Campo-Grosina units and those imaged in the IAM5 section (Fig. 16a) show that the Alpine analogues represent a valuable natural laboratory to study present-day magma-poor rifted margins. Because in the Alps also the age and conditions under which the rift structures form can be determined, these data is also important to constrain and test kinematic and dynamic models of margin formation.

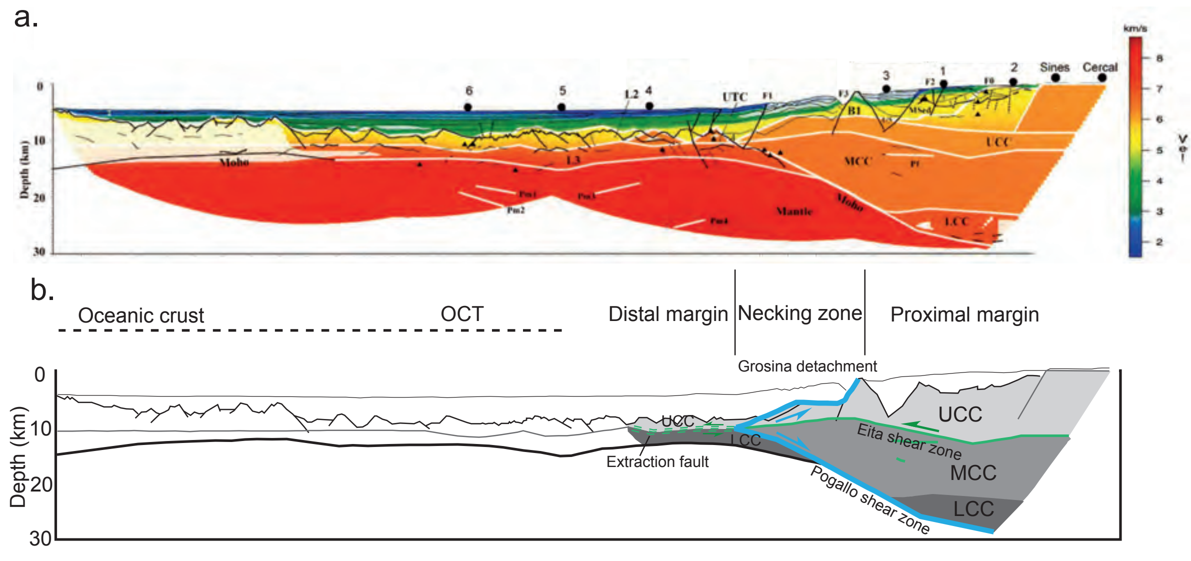


Fig. III. 16: (a) Refraction seismic section across the West Iberia continental margin at 38°N (after Afilhado et al, 2008). (b) Comparison of the crustal structure of the West Iberia continental margin (subdivision in upper, middle and lower crust after Afilhado et al. 2008) and the structures defined in this work.

8. Conclusion

From many present day margins, it could be observed that the major change in crustal thickness from ±30 km in proximal margins to <10 km in hyper extended domain in the distal margin occurs in a narrow zone, which is referred to as the necking zone. In this paper, we have identified and investigated relics of such a necking zone and the adjacent distal margin in the Campo-Grosina and Bernina-Margna units in the Alps. The study of these units enables to identify the crustal architecture of the necking zone and to study the sedimentary and deformation processes associated with extreme crustal thinning during rifting in the Alpine Tethys margins. The main conclusions of this study are:

- The Campo-Grosina necking zone is made by pre-rift upper and mid crustal levels that were juxtaposed and exhumed, as indicated by new thermochronological data within the necking zone during Pliensbachian time (184,55±1,22 Ma and 188,67±0,86 Ma on $^{40}Ar/^{39}Ar$ on biotite) when rifting started to localize within the future distal margin.
- Thinning in the necking zone is the result of the interplay of detachment faulting in the brittle layers (e.g. Grosina detachment in the upper crust and Pogallo-type shear zone in the lower brittle crust) and decoupling and thinning in ductile quartzo-feldspatic middle and lower crust levels along localized ductile decollements (e.g. Eita shear zone)
- The Bernina-Margna units derived from the distal margin, where pre-rift upper crust is juxtaposed against pre-rift lower crust along a major "extraction fault" (e.g. Margna shear zone) responsible for extreme crustal thinning. When the crust is thinned to < 10km,ductile layers become brittle, which enables faults to cut from the surface into mantle and exhume the brittle crust and exhumed mantle at the seafloor (e.g. Bernina and Err detachments) resulting in the delamination of the previously thinning haningwall block (e.g. Briançonnais unit).

These results enable us to propose a coherent structural and tectonic model for the evolution of magma-poor rifted margins during major lithospheric thinning. The necking zone is shown to play a major role in the individualization of the future continents and the formation of a key stone block (also referred to as H-block; Lavier and Manatschal 2006) inbetween (e.g. Briançonnais domain) that will be delaminated during final rifting by low-angle detachment faulting leading finally to the exhumation of crustal and subcontinental mantle in the distal margin and adjacent OCT.

New structural and thermochronological data of the Adriatic rifted margins enable us to propose a coherent evolution for rifting in Alpine Tethys margin in order to explain major crustal thinning. We believe that these results will help to better understand margins undergoing extreme thinning as well as to constrain and test rift models.

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QUATRIÈME PARTIE

Le chapitre trois intègre l'ensemble des données stratigraphiques, structurales, et thermochronologiques accessibles dans le domaine Alpin qui préserve l'enregistrement du rifting. Cette synthèse a pour but l'établissement d'un modèle de l'évolution des marges passives Alpines durant le rifting. La revue des données bibliographiques sur les Alpes est jointe et confrontée aux observations effectuées lors de nombreux stages de terrain dans l'ensemble de la chaîne Alpine concerné par le processus de rifting (bassin de Bourg d'Oisans, Briançonnais, Préalpes Suisses, bassin de Monte Generoso, Canavese, zone d'Ivrée).

Ce chapitre est composé de cinq parties :

La première partie présente brièvement l'histoire de l'avancée des connaissances sur l'ouverture de la Téthys Alpine ainsi qu'une introduction sur le système Alpin.

La deuxième partie décrit l'évolution stratigraphique des différents domaines des marges passives de la Téthys Alpine (marge Européenne, domaine Valaisan, Briançonnais, domaine Liguro-Piémontais, marge Adriatique). Elle conduit à une compilation des caractéristiques stratigraphiques pendant les phases anté-, syn- et post-rift.

La troisième partie est une synthèse de l'histoire tectono-métamorphique et thermochronologique des différents socles présents. L'accent est mis sur les intrusions anté-rift permiennes qui sont des marqueurs traçant l'exhumation des divers niveaux crustaux pendant le rifting. Cette synthèse est couplée aux données des travaux sur le manteau du domaine Alpin.

La quatrième partie est focalisée sur les structures tectoniques extensives qui sont décrites par les travaux antérieurs.

La cinquième partie est une discussion intégrant les données précédentes. A partir de cette synthèse, je propose des modèles évolutifs en 2D et 3D pour contraindre les mécanismes de création de la Téthys Alpine.

Cette section de la Thèse a été publiée dans « International Journal of Earth Sciences » (IJES) pour un volume spécial pour les 100 ans du journal.

Publication 3: Published in IJES for the 100 year anniversary of the GV

UNRAVELLING THE INTERACTION BETWEEN TECTONIC AND SEDIMENTARY PROCESSES DURING LITHOSPHERIC THINNING IN THE ALPINE TETHYS MARGINS

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ABSTRACT

The discovery of exhumed continental mantle and hyper-extended crust in present-day magma-poor rifted margins is at the origin of a paradigm shift within the research field of deep water rifted margins. It opened new questions about the strain history of rifted margins and the nature and composition of sedimentary, crustal and mantle rocks in rifted margins. Thanks to the benefit of more than one century of work in the Alps and access to world-class outcrops preserving the primary relationships between sediments and crustal and mantle rocks from the fossil Alpine Tethys margins, it is possible to link the subsidence history and syn-rift sedimentary evolution with the strain distribution observed in the crust and mantle rocks exposed in the distal rifted margins. In this paper we will focus on the transition from early to late rifting that is associated with considerable crustal thinning and a reorganization of the rift system. Crustal thinning is at the origin of a major change in the style of deformation from high-angle to low-angle normal faulting which controls basin-architecture, sedimentary sources and processes, and the nature of basement rocks exhumed along the detachment faults in the distal margin. Stratigraphic and isotopic ages indicate that this major change occurred in late Sinemurian time, involving a shift of the syn-rift sedimentation towards the distal domain associated with a major reorganization of the crustal structure with exhumation of lower and middle crust. These changes may be triggered by mantle processes, as indicated by the infiltration of MOR-type magmas in the lithospheric mantle, and the uplift of the Briançonnais domain. Thinning and exhumation of the crust and lithosphere also resulted in the creation of new palaeogeographic domains, the Proto-Valais and Liguria-Piemonte domains. These basins show a complex, 3D temporal and spatial evolution that might have evolved, at least in the case of the Liguria-Piemonte basin, in the formation of an embryonic oceanic crust. The re-interpretation of the rift evolution and the architecture of the distal rifted margins in the Alps have important implications for the understanding of rifted margins worldwide, but also for the paleogeographic reconstruction of the Alpine domain and its subsequent Alpine compressional overprint.

Keywords: Magma-poor rifted margins; Alps; Tethys; Thinning; Extensional tectonics

1. Introduction

The discovery of hyper-extended crust in the distal rifted margins of Iberia-Newfoundland (Péron-Pinvidic and Manatschal 2009), of South, Central and North Atlantic (Contrucci et al. 2004; Moulin et al. 2005) and of NW- and S-Australia (Karner and Driscoll 2000) results to new important questions regarding the mechanisms leading to the observed lithospheric thinning and continental break-up and the way these processes are recorded in the stratigraphy. Although lithospheric thinning is a first-order plate tectonic process, at present little is known about how this process develops and results in the formation of deep-water rifted margins. Currently, many projects, mostly in the hydrocarbon industry, aim to map the structure of deep-water rifted margins using high quality reflection and refraction seismic surveys. These studies form the basis for the development of kinematic and dynamic models to investigate the thermo-mechanical evolution of final rifting and continental breakup. However, with the exception of the Iberia-Newfoundland rifted margins, little is known from present-day rifted margins about the nature of the crust and mantle and the structures that accommodate crustal thinning and mantle exhumation. Key questions are how, where and when the lithosphere is thinning and how this process is documented in the geological record. In order to answer these questions it is required to have direct access to the most complete lithospheric sections available, where mantle and deeper crustal rocks, syn-extensional sedimentary record as well as the structures that accommodated crustal thinning can be studied. There is no present-day rifted margin from which we have access to such a complete dataset. A valuable source of critical data, complementary to that of present-day rifted margins, comes from collisional orogens, and in particular from the Alpine Tethys margins preserved in the Alps.

In this paper we will discuss the large scale structures and the types and relative distribution of sedimentary, crustal and mantle rocks from the former Alpine Tethys margins, which are exposed within the Alpine domain of Western Europe. In contrast to previous papers describing the Alpine Tethys margins (e.g. Manatschal 2004 and Manatschal et al. 2007), which reviewed the evolution of the proximal and most distal Alpine Tethys margins, in this paper we focus on extreme crustal thinning that predates mantle exhumation in the Alpine Tethys. At present-day rifted margins, extreme crustal thinning is preserved in the transition zones between the proximal and distal margins, which will be referred to as the "necking zones". The rift evolution of these necking zones and distal domains, which are exposed in the Lower Austroalpine and Penninic Units in the Central Alps, are poorly understood. In the Lower Austroalpine units in Grisons (SE Switzerland), rift-related structures are locally well preserved and the primary relationship between mantle and crustal rocks and the associated stratigraphic record can be observed. In the Penninic Units derived from the conjugate European margin, rift structures were more heavily affected by Alpine deformation. Despite these difficulties, we believe that mapping pre-Alpine rift related field relationships can provide important insights on how, where and when the continental lithosphere is thinning within a conjugate rift system. Since the conjugate rifted margins are exposed in the hanging and footwall of the Alpine subduction zone, the understanding of the sedimentary and crustal architecture of the most distal parts of these rifted margins may also help to better understand how and where subduction initiated and how the rift-inheritance may have controlled the architecture and evolution of the earliest stages of continental collision.

2. Rifted margins in the Alps

2. 1. Historical background

Already in the beginning of the 20th century, Gustav Steinmann noted the close relationship between deep-water sediments and basaltic rocks in the Alps (Steinmann 1925, 1927). This and numerous other pioneering studies (for a review see Bernoulli and Jenkyns 2009) set the base for the present research on rifted margins in the Alps. The discovery of rift-related detachment faults in the most distal rifted margin in the Err nappe in SE Switzerland by Froitzheim and Eberli (1990), together with ODP drilling off Iberia (Boillot et al. 1987, for a review see Tucholke and Sibuet 2007), resulted in a new view of how the lithosphere extends and oceans form. At present there is general agreement on the observation that the architecture of distal and proximal rifted margins are different and that the boundary between continental and oceanic crust is transitional rather than sharp. Despite the significant number of studies describing the structures and processes related to mantle exhumation (Beslier et al. 1995, Froitzheim and Florineth 1994; Manatschal and Nievergelt 1997, Müntener and Manatschal 2006) at present little is known about the mechanisms leading to extreme lithospheric thinning (e.g. Karner and Gambôa 2007, Whitmarsh et al. 2001, Pérez-Guissinyé and Reston 2001, Lavier and Manatschal 2006). If at present-day rifted margins the first order structure as well as the variation of crustal thickness can be determined (e.g. Reston 2009), the very scattered data set, in particular drill hole data, makes it difficult to unravel the precise tectonic and stratigraphic evolution of the lithosphere during extreme crustal thinning.

In this paper we use the example of the Alpine Tethys rifted margins preserved in the Alps, to address three major questions: (1) how and when did the lithosphere thin; (2) what are the structures that document this lithospheric thinning and (3) how are these processes recorded in the structural, stratigraphic and petrographic record.

2. 2. The Alpine Tethys rifted margins

Remnants of the proximal rifted margins are well preserved in the external parts of the Alps (fig 1) and have been extensively described recently by Lemoine et al. (1986), Bertotti et al. (1993), and Manatschal (2004). The major structures forming these proximal margins are fault-bounded basins associated with tilted blocks. They can be mapped over hundreds of kilometers on both sides of the future distal domains. In contrast, the evolution and architecture of the distal margins is much less understood. In present-day magma-poor rifted margins, the distal domains correspond to complex areas that are limited continentwards by necking zones, corresponding to the transition from equilibrated crust, with a thickness of about 30 km, to less than 10 km thick crust over a horizontal distance of less than 50-60 km. This change in crustal thickness is associated with a strong continentward dip of the Moho underneath the necking zone that can be up to 35° (e.g. Lau et al. 2006; Osmundsen and Ebbing 2008; van Avendonk et al. 2009). The distal margin is formed by less than 10km thick that wedges out oceanwards and is replaced by exhumed subcontinental mantle that grades into oceanic crust (for a more detailed description see Péron-Pinvidic and Manatschal 2009). In the Alps, it is impossible to determine crustal thickness at any given time during the rift evolution. However, the occurrence of domains characterized by exhumed continental mantle, referred to as Ocean Continent Transitions (OCT) (e.g. (Err and Platta OCT: Froitzheim and Eberli 1990 and Manatschal and Nievergelt 1997; Tasna OCT: Froitzheim and Florineth 1994 and Manatschal et al. 2007; Margna-Malenco: Hermann and Müntener 1996) clearly demonstrates that within the Alpine domain the crust was locally completely omitted. Although at present it is generally accepted that final mantle exhumation is related to detachment faulting, it is still a matter of debate of how and when these faults became active and how far they can explain the observed extreme crustal thinning (for further discussion see Manatschal 2004 and Reston 2009). Of particular importance are necking zones, which separate the distal and proximal margins and where major thinning is observed in present-day rifted margins. In the Alps, the necking zones of the Adriatic and European rifted margins are preserved in the Lower Austroalpine and North Penninic units, respectively (Fig 1c,d,e). In this paper, we will show that these areas separate domains showing a major change in the stratigraphic and structural evolution as well as in the nature of the basement rocks. These units also separate less deformed "external" from more deformed "internal" Alpine domains.

A long-standing controversy in Alpine geology is the paleogeographic setting of the part of the Western Tethys that was later accreted to the "internal" parts of the Alps, which can be subdivided into the ophiolite-bearing Piemonte-Liguria and Valais domains that surround the Briançonnais domain. The latter represents an enigmatic continental block in the distal domain characterized by a distinctive stratigraphic and isostatic history that enables to distinguish it from the adjacent European and Adria units (Fig 1).

Numerous studies (e.g. Frisch 1979, Stampfli 1993, Wortmann et al. 2001) have addressed the issue of the original position and provenance of the Briançonnais domain. In the past it has been proposed that the Briançonnais domain was attached to Iberian plate and was separated from the Euopean plate by an oceanic domain, the so called "Valais ocean" that was interpreted to represent the eastern prolongation of the Bay of Biscay (Frisch 1979, Stampfli 1993). However, the lack of any evidence for such an oceanic basin in the Western Alps, southern Provence, and eastern Pyrenees makes this interpretation very controversial and poorly supported by observations and data. However, in the Central Alps, it is generally accepted that the Valais domain existed separating the European margin from the Briançonnais domain. The latter represents a portion of continental crust that forms a block in between the European and Adriatic margins. Lavier and Manatschal (2006) interpreted the Briançonnais as a hanging (H)-block, i.e. key stone that formed in between Europe and Adria during the Jurassic rifting (for more discussion see below).

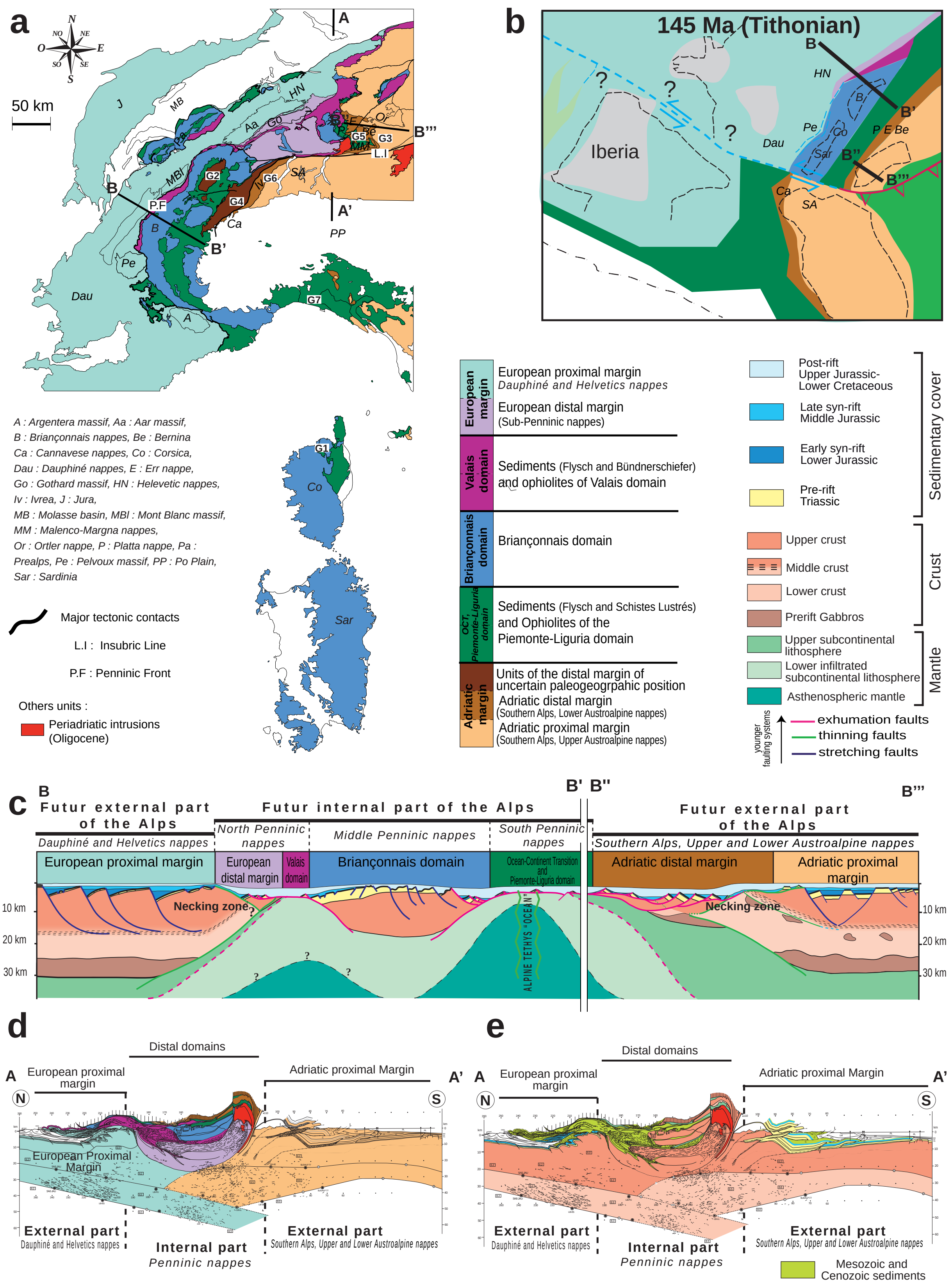
The Valais domain (Trümpy 1980) located between the Briançonnais and European domains (Fig 1b,c) is characterized by remnants of Palaeozoic ophiolites and continental basement units (Beltrando et al. 2007; Masson et al. 2008) associated with Mesozoic pre-, syn- and post-rift sediments that grade upsection into flysch-type deposits recording the onset of Alpine convergence. At present there is a debate as to whether this domain represents a branch of an Early Cretaceous ocean, as suggested by Frisch 1979, Trümpy 1980, Stampfli 1993, or whether it represents a Jurassic basin that was locally floored by exhumed mantle (Manatschal and Müntener 2009). The problem of restoring and interpreting the Valais domain is partly a consequence of the strong Alpine overprint, and also of the complex pre-Alpine architecture. Future debates about nature, age and paleogeographic position of the Valais domain will need to integrate more realistic margin architectures as well as the stratigraphic record documented within this basin.

The Liguria Piemonte domain located between the Briançonnais and Adriatic domains (Fig 1b, c) has been largely subducted and only relics of this domain are preserved as ophiolites in the Alps. While in the western Alps most of these remnants show a high pressure metamorphic overprint, in the eastern Alps in Grisons, these units are less affected by Alpine deformation. In the Lower Austroalpine units in Grisons (SE Switzerland) remnants of its southeastern margin are spectacularly exposed and provide access to different crustal levels and complex sedimentary and deformation systems that are related to the Mesozoic thinning of the continental crust. In this paper, we will focus mainly to this section, since it provides the best record of how the crust was thinning during Jurassic rifting.

3. Tectono-sedimentary evolution of the Alpine margins

3. 1. Pre-rift evolution of the Alpine domain

The pre-rift stratigraphic record of the Alpine Tethys rift system starts in late Carboniferous to early Permian time documenting the final convergence and orogenic collapse of the Variscan orogenic system. By the end of Permian to early-middle Triassic time, most of the topography related to the Variscan orogen disappeared and a sedimentary cycle initiated with a widespread siliciclastic sequence derived from the erosion of topographic highs (e.g. Verrucano and Bundsandstein Formations). Up-section, these continental deposits are replaced by shallow marine platform sediments that are the result of a major transgressive system. In detail the evolution is more complex (Bechstädt et al. 1978; Channel and Kosur 1997; Robertson 2004), since it was associated with


a
50 km
A : Argentera massif, Aa : Aar massif,
B : Briançonnais nappes, Be : Bernina
Ca : Cannavese nappes, Co : Corsica,
Dau : Dauphiné nappes, E : Err nappe,
Go : Gothard massif, HN : Helevetic nappes,
Iv : Ivrea, J : Jura,
MB : Molasse basin, MBl : Mont Blanc massif,
MM : Malenco-Margna nappes,
Or : Ortler nappe, P : Platta nappe, Pa :
Prealps, Pe : Pelvoux massif, PP : Po Plain,
Sar : Sardinia
Major tectonic contacts
L.I : Insubric Line
P.F : Penninic Front
Others units :
Periadriatic intrusions (Oligocene)
b
145 Ma (Tithonian)
Iberia
European proximal margin
Dauphiné and Helvetics nappes
European distal margin
(Sub-Penninic nappes)
Sediments (Flysch and Bündnerschiefer) and ophiolites of Valais domain
Briançonnais domain
Sediments (Flysch and Schistes Lustrés) and Ophiolites of the Piemonte-Liguria domain
Units of the distal margin of uncertain paleogeogrpahic position
Adriatic distal margin
(Southern Alps, Lower Austroalpine nappes)
Adriatic proximal margin
(Southern Alps, Upper Austroalpine nappes)
Post-rift Upper Jurassic-Lower Cretaceous
Late syn-rift Middle Jurassic
Early syn-rift Lower Jurassic
Pre-rift Triassic
Sedimentary cover
Upper crust
Middle crust
Lower crust
Prerift Gabbros
Crust
Upper subcontinental lithosphere
Lower infiltrated subcontinental lithosphere
Asthenospheric mantle
Mantle
younger faulting systems
exhumation faults
thinning faults
stretching faults
c
Futur external part of the Alps
Dauphiné and Helvetics nappes
Futur internal part of the Alps
North Penninic nappes
Middle Penninic nappes
South Penninic nappes
Futur external part of the Alps
Southern Alps, Upper and Lower Austroalpine nappes
European proximal margin
European distal margin
Valais domain
Briançonnais domain
Ocean-Continent Transition and Piemonte-Liguria domain
Adriatic distal margin
Adriatic proximal margin
Necking zone
ALPINE TETHYS "OCEAN"
10 km
20 km
30 km
d
Distal domains
European proximal margin
Adriatic proximal Margin
European Proximal Margin
External part
Dauphiné and Helvetics nappes
Internal part
Penninic nappes
External part
Southern Alps, Upper and Lower Austroalpine nappes
e
Mesozoic and Cenozoic sediments

continental rifts and climate changes that may explain local change in sedimentary thicknesses and incursions of evaporitic systems between the future European, Briançonnais and Adriatic domains. The transgression within the future European domain occurred only in Late Ladinian or Early Carnian time and was accompanied by little subsidence, resulting in a thin, only a few tens of meters thick sequence formed by limestone, dolomite, shale and evaporites (Muschelkalk and Keuper) (Fig. 2a). In contrast the transgression onto the future Briançonnais and Adriatic domains started earlier and was accompanied by major subsidence, as indicated by several kilometer thick carbonate platforms. In the future Briançonnais and Adriatic domains two major carbonate platforms can be distinguished, a Ladinian and a Norian one, intercalated by a major evaporitic system corresponding to the Carnian Raibl Formation (Bosellini 1984). These carbonate platform deposits overlie continental conglomerates and sandstones of Upper Permian to Lower Triassic age referred to as Verrucano formation s.l. (e.g. Assereto et al. 1973). The thickening of the Triassic carbonate sequences to the east/southeast and their transition into deep marine facies suggest that these deposits may have formed as syn- to post-rift sedimentary sequences related to rifting and opening of the Meliata/Vardar oceanic domains along the eastern border of Adria (Dercour et al. 2000) (Fig. 2a). Thus, these units cannot be called “pre-rift” sensu strictu, although, with respect to the Alpine Tethys rift system here referred to as the rifting event that resulted in the separation of Adria from Europe/Briançonnais, the Triassic sequences are clearly “pre-rift”.

3. 2. Syn-rift evolution of the Alpine Tethys domain

The timing of the onset of rifting in the Alpine Tethys domain is difficult to determine, due to the superposition of different rift-systems. However, at the end of Triassic to earliest Jurassic time, rifting migrated into the area that eventually became the Alpine Tethys domain (Fig. 2b). From Hettangian time onwards fault-bounded rift basins formed across the entire future Adriatic and European proximal margins (Lemoine et al. 1993, Eberli. 1988; Bertotti et al. 1993). This extensional activity continued until Toarcian to Aalenian time (Lemoine and Trümpy 1986 and Bertotti et al. 1993). During this stage, water depth in the basins never exceeded 1000 m (Roux et al. 1988), whereas the swells remained shallow, Subaerial exposure was locally attained, as indicated by erosion of the upper Triassic sediments in the footwall block and their re-deposition in the basin in early Hettangian (e.g. Il Motto; Eberli 1988). As a consequence, a large variety of sedimentary facies can be found, ranging from me-

Fig. IV. 1: Maps (a, b) and cross sections (c, d, e) showing the distribution of major Alpine tectonic units and their paleogeographic position during Late Jurassic time. (a) Tectonic map of the Central and Western Alps including Corsica, Sardinia and the northern Apennine (modified from Schmid et al 2004). For major paleogeographic units distinguished in the maps and sections and locations referred to in the text see legend in the map. (b) Paleogeographic map of the Alpine domain showing the situation at the end of Jurassic (Tithonian: 145 Ma) (modified after Leleu et al. submitted). (c) Cross section through the Alpine Tethys showing the Late Jurassic crustal structure of the rifted margin. Note that the section across the Adriatic rifted margin (section across the Austroalpine and South Penninic units in Grisons) is offset from the section across the conjugate European rifted margin (section across the Dauphine – Briançonnais – Valais units along the French-Italian border). (d and e) Interpreted NFP 20 seismic section across the Central Alps (from Schmid et al. 1996) showing the distribution of the paleogeographic units (d) and the various lithologies (e). For nature of lithologies see legend above.

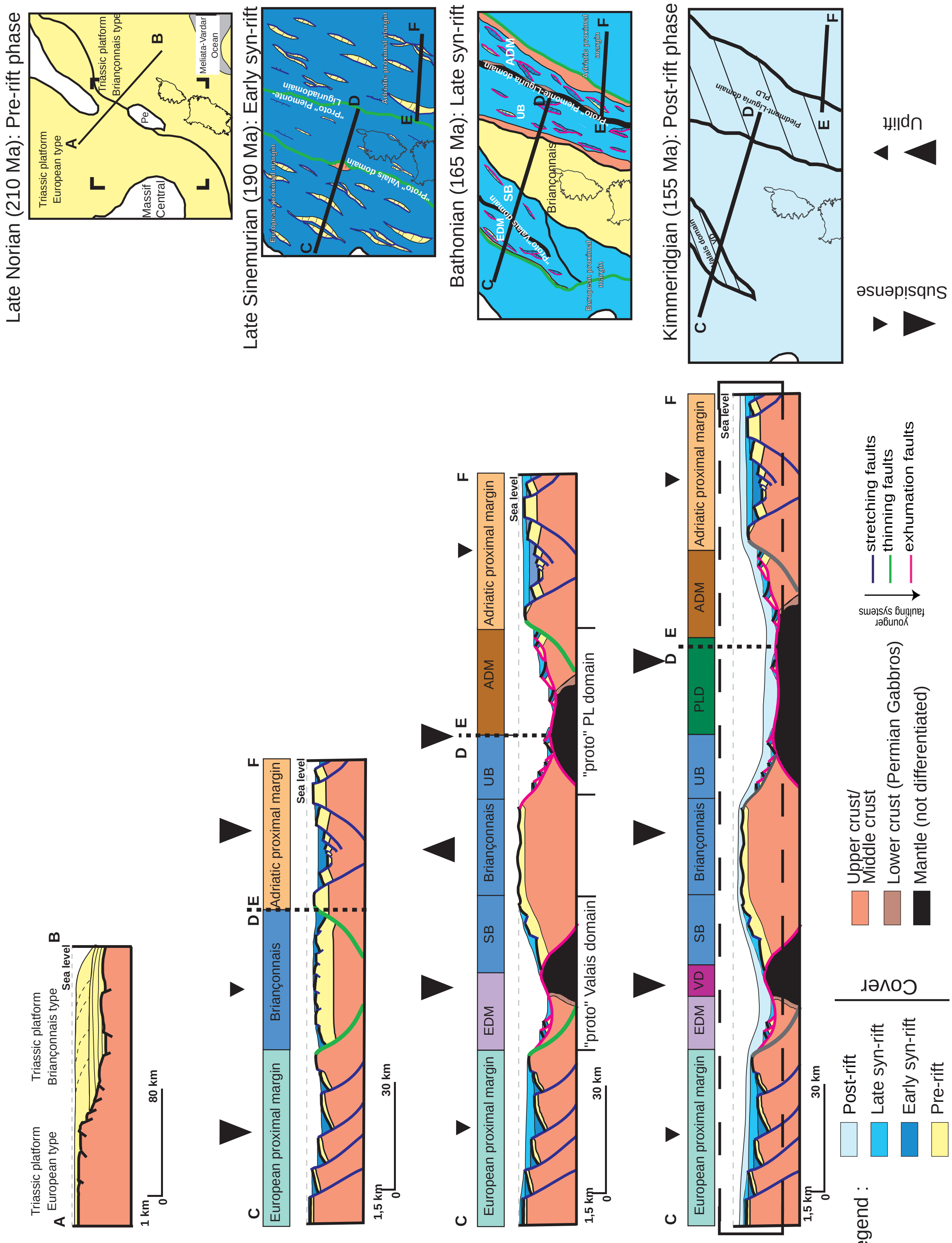

Late Norian (210 Ma): Pre-rift phase
Triassic platform European type
Triassic platform Briançonnais type
Massif Central
Meliata-Vardar Ocean
Pe
A
B
Sea level
1 km
0
80 km
Late Sinemurian (190 Ma): Early syn-rift
"proto" Valais domain
"proto" Piemonte-Liguria domain
European proximal margin
Adriatic proximal margin
C
D
E
F
Briançonnais
D E
1,5 km
0
30 km
Bathonian (165 Ma): Late syn-rift
EDM
SB
UB
ADM
European proximal margin
Briançonnais
"proto" Valais domain
"proto" PL domain
Kimmeridgian (155 Ma): Post-rift phase
Piedmont-Liguria domain
PLD
VD
Subsidence
Uplift
Legend :
Cover
Post-rift
Late syn-rift
Early syn-rift
Pre-rift
Upper crust/ Middle crust
Lower crust (Permian Gabbros)
Mantle (not differentiated)
younger faulting systems
stretching faults
thinning faults
exhumation faults

Fig. 3

gabreccias and olistoliths close to fault scarps, to calc-turbidites intercalated with hemipelagic sediment in more distal parts of the basins, to condensed sediments with stratigraphic gaps on the swells (Bertotti et al. 1993, Eberli 1988) (Figs. 2b and 3). The sediments are mainly derived from the widespread Triassic carbonate platforms with only subordinate input of continental material.

From Late Sinemurian time onwards the basins in the proximal parts of the future margins were sealed and extension became more localized within the future distal margin (Eberli 1988, Bertotti et al. 1993, Berra et al. 2009) (Figs. 2c and 3). This major change corresponds to the onset of a rift phase, which led to major crustal thinning and to the formation of small embryonic oceanic domains within the Alpine realm (Figs. 2b and c). Due to the stronger Alpine tectonic overprint, the more internal domains and in particular those that become part of the subducting plate, i.e. those associated with the European margin, are difficult to restore and interpret. However, three paleogeographic domains, the Liguria-Piemonte, Briançonnais and Valais domains corresponding to the Upper-, Middle- and Lower-Penninic Alpine tectonic units (Schmid et al. 2004 and references therein) individualized during this stage of rifting. These domains/units can be traced along most of the Alpine chain and are characterized by a distinct stratigraphic record and a tectonic position within the former margin. All three major paleogeographic domains underwent crustal thinning, but only the Liguria-Piemonte and locally also the Valais domains, developed into embryonic oceanic domains, i.e. domains that were floored by subcontinental mantle and were eventually affected by subordinate MOR intrusive and extrusive magmatic activity (e.g. Manatschal and Müntener 2009). Since the terms “Liguria-Piemonte” and “Valais” are commonly used to describe either an oceanic domain or ophiolites, we introduce in this paper the term “proto” Liguria-Piemonte and Valais domains to describe in the following the Late Sinemurian to late middle Jurassic paleogeographic evolution of these domains/basins pre-dating the onset of mantle exhumation (Fig. 2c).

Fig. IV. 2: Reconstructed sections and paleogeographic maps illustrating the paleogeographic evolution of the Alpine Tethys margins showing the situations in the: (a) Norian (pre-rift): transgressive shallow marine carbonate platforms; (b) Sinemurian (early syn-rift): high-angle normal faulting and tilted blocks occurring across the whole future margin. (c) Bathonian (late syn-rift): thinning and exhumation in the Proto Valais and Proto Piedmont-Liguria basin that are separated by the uplifted Briançonnais domain. Basins in the proximal rifted margins are sealed and are not anymore tectonically active; (d) Kimmeridgian (post-rift): passive infill and onlapping onto inherited topography (onlapping towards the necking zone). In contrast to the Valais domain, the Piemonte-Liguria domain continues to extend after first mantle exhumation as documented by the formation of an embryonic oceanic crust. Norian paleogeographic map is redrawn after Dercour et al. 2000

3. 3. Briançonnais domain

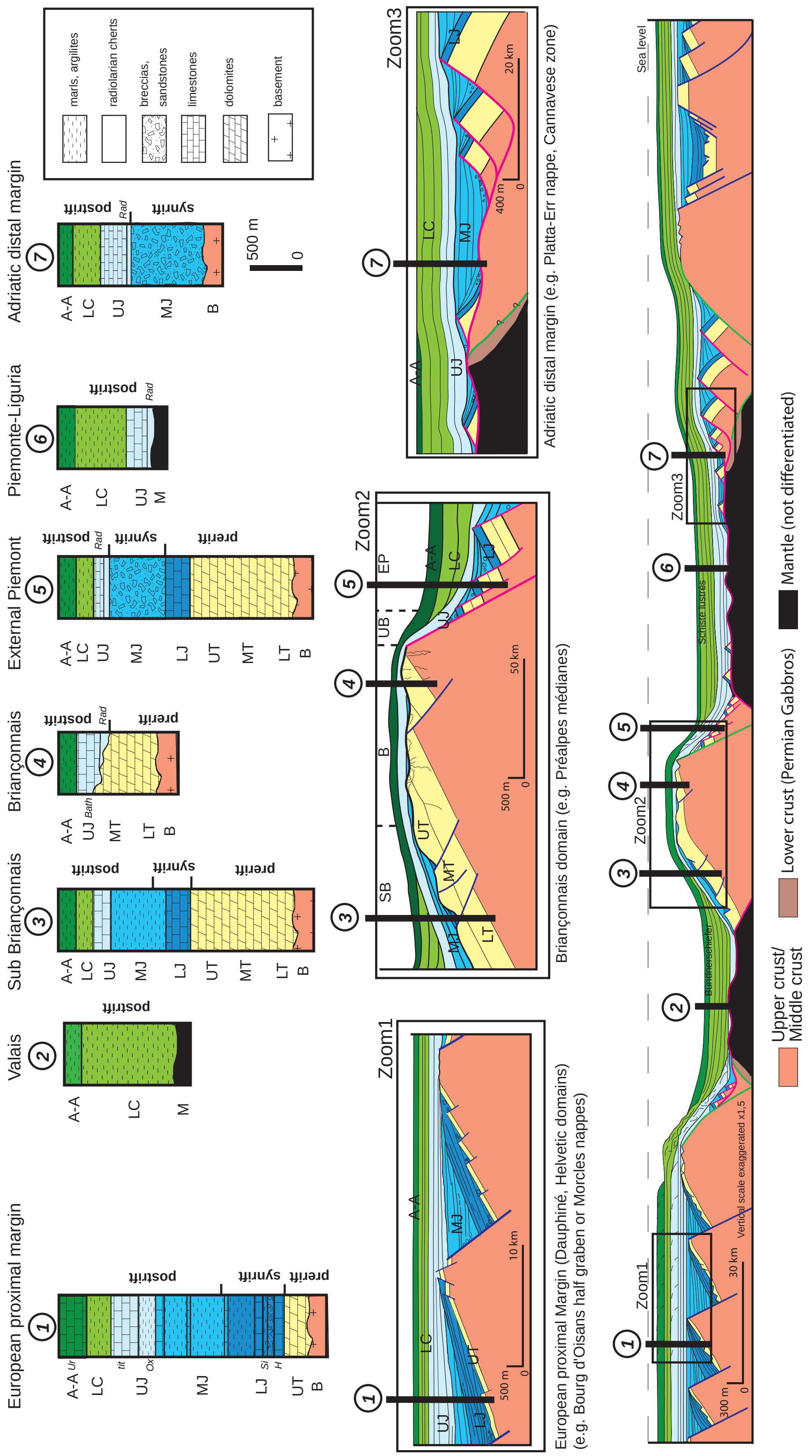
The Late Permian to Early Liassic stratigraphic evolution of the Briançonnais domain shows many similarities with the one observed in the future proximal Adriatic margin. The early Liassic consists of mainly neritic limestones with rare normal faults associated with the formation of small shallow rift-basins never exceeding 200 to 600 m water depth (Roux et al. 1988, Mettraux and Mosar 1989, Claudel and Dumont 1999). After initial subsidence, the Briançonnais domain can be subdivided in three main paleogeographic areas: (1) the internal Briançonnais and the (2) northwestern and (3) southeastern margins identified as Subbriançonnais and Ultrabriançonnais respectively. From Toarcian onwards the internal part of the Briançonnais emerged and was affected by a karstic environment (Debelmas 1955, 1987, Lemoine et al. 1986, Faure and Mégard-Galli. 1988, Ellenberger 1955, Jaillard 1985, Bloch 1963, Vanossi 1965, Baud and Masson 1975, Baud et al. 1979). Evidence for sub-aerial exposure and partial erosion of the Triassic and Lower Jurassic platform can be found across the entire internal Briançonnais domain (Claudel and Dumont 1999) indicating regional rather than local uplift. The karst affects the uppermost 300 m of the emerged carbonate platform (Faure and Mégard-Galli 1988) and is associated with continental deposits that fill the karst and locally unconformably overlie the emerged platform. The sediments are made by calcareous breccias with a vadose cementation, conglomerates with Triassic and Liassic pebbles and paleosoils. In-situ brecciation and infill of fractures are also observed (Faure and Mégard-Galli 1988, Claudel and Dumont 1999). The emersion and the related erosion hinder any detailed reconstruction of the evolution of this domain during initial rifting. The major hiatus in sedimentation that is related to this uplift is a characteristic feature for the entire internal Briançonnais domain. After emersion, the Briançonnais domain subsided rapidly, as indicated by the sealing of the karst by open marine neritic to pelagic sediments.

3. 4. Proto Piemonte-Liguria and Proto Valais basins

Deep-water, hyper-extended basins developed between the Briançonnais and the future Adriatic and European proximal margins from Sinemurian time onwards (Fig. 2c and 3). Extension in these basins overprinted the previous Early Jurassic rift structures and resulted in the final exhumation of mantle rocks in late Middle Jurassic time (Bathonian to Callovian). The structures that document final mantle exhumation in these basins are well-described for the Piemonte Liguria domain but are much more controversial for the Valais domain (see Manatschal and Müntener 2009).

The stratigraphic and tectonic evolution of the distal Adriatic margin is well preserved in the Bernina and Err nappes in SE Switzerland and in the westernmost Southern Alps (Canavese, Biella, Cusio zone) in NW Italy (Fig. 1). In these zones the Triassic and Lower Jurassic carbonates are locally absent, a feature that may either be explained by emersion (non deposition: Berra et al. 2009), uplift and erosion during early Jurassic (tectonic uplift; e.g. Lemoine et al. 1987), or tectonic omission due to exhumation (see discussion below). Pliensbachian and younger sediments overlie unconformably basement rocks or sediments ranging in age from lower Jurassic to Triassic (e.g. Manatschal and Nievergelt 1997, Ferrando et al. 2004, Berra et al. 2009) (Fig. 3). This is best shown in the Err nappe, where the top of the pre-rift sequence (e.g. Agnelli formation) is made by a hardground that has been dated by ammonites to the Late Pliensbachian (Furrer et al. 1985). This hardground is unconformably overlain by carbonate-dominated debris flows and related finer deposits (e.g. Bardella formation) that develop into mainly siliciclastic sequences upsection (e.g. Saluver formation; Finger 1978) (Fig. 3). The change from mainly carbonates to siliciclastic sediments during final rifting reflects a major change in the source area that is related, as discussed below, to the exhumation of crustal rocks along exhumation faults in the Err/Bernina domain (e.g. Fig. 5). The occurrence of Lower Cretaceous sediments directly overlying basement rocks in the Bernina domain (e.g. Piz Corvatsch Spillmann 1993, Peters 2005) indicates that the basins were underfilled and highs remained emerged at deep water depth at least until Aptian/Albian time. Similar relationships to those described from the Bernina nappe were also reported from the Biella and Cusio zones in the western Southern Alps (Berra et al. 2009). The Err and Canavese domains may represent more thinned continental crust, as indicated by their more distal position and their proximity to exhumed mantle domains (Figs. 1 and 3). A more distal position is also indicated by the stratigraphic record that shows a starved but continuous evolution from syn-rift into post-rift sediments (Ferrando et al. 2004).

The northwestern margin of the Proto Piemonte-Liguria basin is preserved in the Ultrabriançonnais and external Liguro-Piemontais domains, which are best preserved in South-Eastern France. These domains mark the transition between the Briançonnais and Liguro-Piemonte embryonic oceanic domain. The Ultrabriançonnais is formed by breccias made of basement and dolomite clasts, with a matrix constituted of arkose laying on Paleozoic basement or Late Permian to Early Triassic sediments (e.g. Breccia nappe, Dall'Agnolo 2000). The exact age of the breccias is still debated but they are interpreted to be Middle Jurassic, i.e. related to rifting in the Alpine domain (Dall'Agnolo 2000 and references therein). In the external Liguro-Piemontais domain, which is in a more distal position, the late Early to early Middle Jurassic erosional event observed over the Briançonnais domain is absent and a complete Triassic section is overlain by cherty limestones of Sinemurian to Toarcian age. This succession is followed by breccias and sandstone, which have dolomitic and basement clasts as well as ophiolitic components (e.g. Rochebrune massif: Dumont 1984). The age of this breccia is not well known but the occurrence of ophiolite clasts and the observation that it is sealed by Callovian-Oxfordian radiolarian cherts makes that is has to be latest Early to early Middle Jurassic, i.e. simultaneous with continental breakup. The occurrence of breccias within the Ultrabriançonnais and external Liguro-Piemontais domains was classically interpreted as the result of tectonic activity along high-angle normal faults (Lemoine et al. 1986).


European proximal margin
Valais
Sub Briançonnais
Briançonnais
External Piemont
Piemonte-Liguria
Adriatic distal margin
prerift
synrift
postrift
Rad
marls, argilites
radiolarian cherts
breccias, sandstones
limestones
dolomites
basement
500 m
0
Zoom1
Zoom2
Zoom3
European proximal Margin (Dauphiné, Helvetic domains)
(e.g. Bourg d'Oisans half graben or Morcles nappes)
Briançonnais domain (e.g. Préalpes médianes)
Adriatic distal margin (e.g. Platta-Err nappe, Cannavese zone)
10 km
50 km
20 km
400 m
Sea level
30 km
300 m
Vertical scale exaggerated x1.5
Bündnerschiefer
Schiste lustrés
Upper crust/
Middle crust
Lower crust (Permian Gabbros)
Mantle (not differentiated)

In contrast to the Proto Liguria-Piemonte domain, the evolution of the Proto Valais domain is much less constrained. Only few sections exist that are not too heavily overprinted by Alpine deformation (e.g. Schams and Falknis units in Grisons and in the Prealpes in Western Switzerland (e.g. Allemann 1952, Baud and Steptfontaine 1980, Stampfli 1993)). The stratigraphic evolution of this domain is best documented in the Subbriançonnais units, which were located on the northwestern slope of the Briançonnais domain (Fig. 3, Zoom 2). In the Subbriançonnais an extensional event is well documented during late Early Jurassic and late Middle Jurassic time (Masson and Septfontaine 1979). After an initial uplift, from Aalenian onwards, the Sub Briançonnais was affected by strong subsidence (Baud and Septfonatine 1980; Borel 1995), which was interrupted by a small episode of uplift during Callovian time in the Northern part (Prealps). However, during late Jurassic time this domain remained in a deep water environment that was controlled by the sedimentation of limestones and shales. Much less is known from the European distal margin, which is due to the fact that the sediments documenting the evolution of this domain were either subducted or were accreted and strongly overprinted during Alpine orogenesis. Breccias overlying basement rocks have been observed from the North Penninic units (Masson, pers. comm.). However, due to the strong Alpine tectonic overprint, the age of these breccias is difficult to determine.

3. 5. Post-rift evolution

During Late Jurassic and Lower Cretaceous time, the European margin was formed by a carbonate shelf. The depositional succession is controlled by a prograding carbonate platform (e.g Tithonian and Urgonian platforms, Fig. 3) that is time equivalent to more marly and shaly sediments in more distal parts of the margin. The evolution of the conjugate Adriatic margin is different. The first post-rift deposits are radiolarian cherts deposited in Bathonian to Kimmeridgian time that are overlain by micritic limestones (e.g. Aptychus limestone/Maiolica; Weissert and Bernoulli 1985) and shales (e.g. Emmat serie, Finger 1978). These facies can be mapped from distal towards more proximal parts of the margin. In between the Adriatic and Europeans margins, a complex domain made respectively from NW to SE by: (1) Valais basin, (2) Briançonnais domain and finally the (3) Liguro-Piemonte basin. During the Late Jurassic the internal parts of the Briançonnais domain are covered by shallow water to open marine neritic limestones (e.g. Prealps, Baud and Septfonatine 1980) to the north, while this domain is overlain by pelagic limestones (e.g. Briançonnais of Briançon, Bourbon 1980) to the south. Early Cretaceous deposits are characterized by highly condensed or absent sedimentation, which persists until Late Cretaceous.

Fig. IV. 3: Section across the Alpine Tethys margin (same trace like sections B-B'" shown in Fig. 1c) showing the stratigraphic record documented in the major paleogeographic domains. Zooms show the detailed sediment-architecture of the proximal European margin (modified from Lemoine et al. 1986), the valais (modified from Steinmann 1994), the Briançonnais (modified after Baud and Stepfontaine. 1980) and the distal Adriatic margin (this work). The simplified composite stratigraphic logs above show typical logs for the different paleogeographic domains. Abbreviations: M, Mantle, B basement. LT, MT, UT, Lower, Middle, Upper Triassic. LJ, MJ, UJ, Lower, Middle, Upper Jurassic; (H, Hettangian, Si, Sinemurian, Bath, Bathonian, Ox, Oxfordian, Tit, Tithonian), LC, Lower Creatceous. A-A, Aptian-Albian (Ur, Urgonian).

While in the southern parts this may be explained by sediment starvation, in the north it can be observed that the Late Jurassic limestone bears a paleokarst sealed by Late Cretaceous pelagic marls (e.g. Sulzfluh nappe, Allemann 1952). This indicates that this domain may have undergone a complex subsidence history, with local areas remaining very shallow even after continental breakup. The different sedimentary record in the Liguro-Piemonte and Valaisan basins, as well as post-rift succession of the Briançonnais domain, indicate that at least from the Late Jurassic onwards two major basins can be distinguished. Both basins contain ophiolitic basement and show evidence for extrusive magmatism (Fig 2d). The Piemonte Liguria basin is dominated by pelagic sedimentation (e.g. Radiolarian Cherts and Aptychus Limestone) while the Valais basin by detrital sedimentation (e.g Terres Noires) associated with deposition of breccias in the Sion-Courmayeur area (Trümpy 1952) may be related with either tectonic or gratitational processes.

In latest Jurassic and Early Cretaceous argillaceous and marly shales were deposited in the two basins (e.g Bündnerschiefer in the Valais and Schistes Lustrés/Argille a Palombini in the Piemonte Liguria basins). The exact age of deposition as well as their tectonic significance is still debated, due to the strong Alpine overprint. However, the observation that Lower Cretaceous sediments onlap onto basement at the margins/necking zones surrounding these basins suggests that these basins remained underfilled until the onset of compression in Campanian time. The post-rift sedimentary sequences developed up-section into classical "flysch" sequences that document the reactivation of the former rifted margins. Due to the more internal position of the Piemonte Liguria basin, the "flysch" sequences in this basin are older (Late Cretaceous to Eocene) than those in the more external Valais basin (Maastrichtian) (Stampfli et al. 1998).

4. Nature and evolution of the crustal and mantle rocks in the Alpine Tethys domain

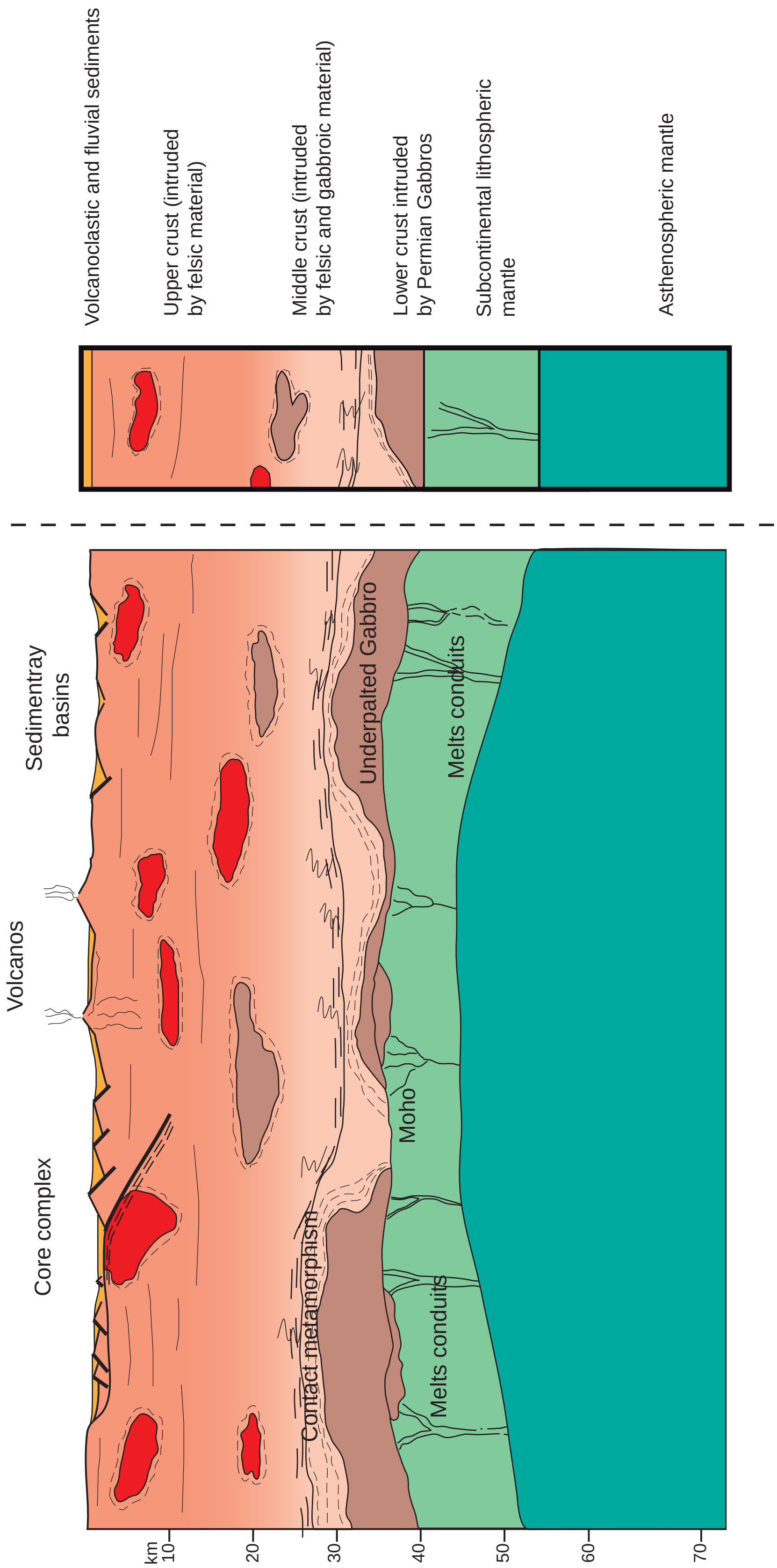
The discovery of extreme crustal thinning and mantle exhumation in distal magma-poor rifted margins opened new questions about the nature, composition and strain history of crustal and mantle rocks within the community working on rifted margins. However, basement rocks from deep water, hyper-extended rifted margins are only known from the Iberia-Newfoundland margins (Tucholke and Sibuet 2007 and references therein). The access to world-class outcrops preserving crustal and mantle rocks from the fossil rifted margins in the Alps became therefore a key to understand how crustal and mantle rocks behave during extreme crustal thinning and mantle exhumation. A pre-requisite to study this evolution is to define the pre-rift history of the crust and mantle (Fig. 4).

4. 1. Continental crust

The present-day continental crust in the Alpine domain has undergone a complex multi-stage tectono-metamorphic history that initiated with the Caledonian orogeny (von Raumer et al. 1998), subsequently overprinted by the Variscan (e.g. Handy et al. 1999) and Alpine (e.g. Beltrando et al.in press) orogenic cycles. It is now widely accepted that the extensional collapse of the Variscan orogen during the Permian and the subsequent Jurassic rifting (e.g. Handy et al. 1999; Schuster and Stüwe 2008 Müntener and Hermann 2001) played a significant role in overprinting previous structures/mineral assemblages of the older continental crust. In this section we will mainly focus on the Late Carboniferous to Early Permian evolution, which significantly pre-structured the crust/lithosphere before it was extended and thinned during Jurassic rifting. Having constrains from the pre-rift crustal structure of the areas that later became part of the rifted margins is a fundamental step to estimate the amount of crustal thinning achieved in different parts of the necking zones and to understand the mechanisms associated with crustal extension. A review of the Permian depth estimates available for representative crustal units of the Alps is presented in Table 1.

During the late Carboniferous to early Permian collapse of the overthickened Variscan crust, the continental crust was strongly modified by i) magma underplating, which produced a gabbroic lower crust and induced a granulitic/amphibolitic metamorphic overprint in the surrounding lower and middle crustal rocks (e.g. Müntener and Hermann 2001, Schuster and Stüwe 2008), ii) emplacement of intrusive felsic plutons at all crustal levels and extrusion of rhyolites at the surface (e.g. Schaltegger and Brack 2007, Dallagiovanna et al. 2009 and references therein) and iii) formation of Permian basins (e.g. Colio basin, Southern Alps (Froitzheim et al. 2008). The crustal thickness at the end of this event was close to 30 km, as indicated by thermobarometric studies (Henk et al. 1997) and by the transgression of shallow marine shelf seas and the rather thin lower Triassic deposits, indicating that the crust was in isostatic equilibrium.

Petrological studies of contact metamorphic aureoles surrounding the Permian intrusives offer the opportunity to constrain the depth at which specific crustal sections were located prior to Mesozoic rifting (see Table 1). This method avoids the potentially misleading attribution of rock types to specific depths exclusively based on their metamorphic grade, since in some localities high-grade rocks of pre-Permian age had been already exhumed to higher crustal levels prior to the widespread late-Variscan tectonometamorphic event. This set of information is of major importance to understand the subsequent deformation history and vertical motions during rifting. The three major crustal layers in the pre-rift crust are (e.g. Fig. 4):


Volcanoclastic and fluvial sediments
Upper crust (intruded by felsic material)
Middle crust (intruded by felsic and gabbroic material)
Lower crust intruded by Permian Gabbros
Subcontinental lithospheric mantle
Asthenospheric mantle
Sedimentray basins
Volcanos
Core complex
Underpalted Gabbro
Melts conduits
Moho
Contact metamorphism
Melts conduits
km
10
20
30
40
50
60
70

(1) *Pre-rift upper crust* consisting of polymetamorphic basement formed locally by high-grade metamorphic rocks. These rocks were already residing at upper crustal level prior to Mesozoic rifting. Only a few Permian gabbros intruded in the pre-rift upper crustal levels (Bocca di Tenda gabbros (Table 1), Tribuzio et al. 2009). More commonly, the late-Carboniferous to early-Permian upper crustal intrusions are felsic and led to the formation of granitoid bodies (von Quadt et al. 1994 Bertrand et al. 2005, Pinarelli et al. 1988; Boriani et al. 1995, Vavra et al. 1999) which are related to acidic to intermediate volcanism at the surface (e.g. Schaltegger and Brack 2007; Dallagiovanna et al. 2009).

(2) Pre-rift middle crust representing the transition between gabbroic intrusions in the lower crust and granitic intrusions in the upper crust. The Mont Collon (Table 1) intrusion in the Dent Blanche nappe of the Western Alps, and the Sondalo gabbro (Table 1) in the Austroalpine units of the Eastern Alps were intruded at mid-crustal depth (e.g. Tribuzio et al. 1999). The metamorphism in the country rocks is characterized by upper greenschist to amphibolite facies mineral assemblages, while narrow granulite facies metamorphism is locally observed in close proximity of the gabbroic intrusions (Braga et al. 2001, 2003).

(3) Pre-rift lower crust consisting mainly of high-grade metamorphic rocks (amphibolite to granulite facies) intruded predominantly by mafic complexes. The Mafic Formation of the Ivrea Zone, in the Southern Alps, represents the most widely studied example of such magmatism. This composite mafic pluton intruded the lower Ivrea crust at 288±4 Ma (Peressini et al. 2007), leading to widespread metamorphic re-equilibration at pressures exceeding 0.8 GPa of the metapelitic rocks (Henk et al. 1997). The Braccia gabbro, in the Malenco unit (Table 1) was intruded at the boundary between lower crust and upper mantle at ca. 270 Ma. From these rocks Müntener et al. (2000) determined the P-T conditions at the crust-mantle boundary at P= 0.8±0.1 Gpa, T= 600±50°C, which represent typical values for thermally equilibrated 30 km thick crust (Fig. 5d). The observed isobaric cooling of the Braccia body, which took place over 50 Myr, indicates that it resided in the lower crust long after its emplacement, before undergoing rapid exhumation near the seafloor during Jurassic rifting.

Fig. IV. 4: Schematic cross-section at the end of Late Permian extension showing the aquired structure of the crust before onset of rifting. Note the stratification of the crust with felsic material preferentially emplaced in upper crustal levels and mafic material in lower crust levels. Note also the granulite facies contact metamorphism of Permian age that is associated with the emplacement of the gabbros and the shallow position of the asthenosphere that is the supposed source for the Permian magmatic rocks emplaced at the base of the crust. To the right there is a simplified section across the pre-rift lithosphere that shows the major crustal levels that constitute the lithosphere before onset of rifting.

Symbole	Location	Lithologies	Lithologies of country rock	P-T Condition of emplacement and re-equilibration	Crystallization Age	Exhumation Age ***Cr*** **: Crystallization age** ***Co*** **: Cooling age**	Evidence of exhumation as reveled by the Sedimentary cover	References
G1	Bocca di Tenda (Northern Corsica)	Olivine-gabbro-norites	Plutonic suite made of Hornblende-granitoïds	Upper Crust Condition of intrusion: ~0,2 GPa	280-290 Ma U-Pb on Zircon	No available data	Primary contact with Permian to Mesozoic rock	Cocherie et al., 2005 Paquette et al., 2003 Poitrasson et al., 1994 Renna et al., 2007 Tribuzio et al., 2009
G2	Mont Collon gabbro (Dent Blanche nappe)	Olivine-gabbro and wehrlites	?High-grade paragneisses and fine-grained metabasites intruded by late Paleozoic granitoids (Arolla Series)?	Middel crust Condition of intrusion: 0,5 GPa-1100°C	284±0,6 Ma U/Pb on Zircon	No data	Polymictic breccias (presumably of Middle Jurassic age) are made by clasts of Arolla series.	Dal Piaz et al., 1977 Ayrton et al., 1982 Monjoie et al., 2005 Monjoie et al., 2007
G3	Sondalo (Austroalpine nappe)	Gabbro-norites, troctolite	Two micas metapelite (<0,9 Gpa , >550°C)	Middle crust Condition of intrusion: 0,3-08 GPa, ~900°C	300-270 Ma Sm-Nd mineral and Rb-Sr on plagioclase-amphibole-whole rock isochron	188±4 Ma $^{40}Ar/^{39}Ar$ on Muscovite, ***Co*** 182±6 Ma K/Ar on Muscovite, ***Co*** 164±7 Ma K/Ar on Biotite, ***Co***	No data	Braga et al., 2001 Braga et al., 2003 Phd of Meier., 2003 Thöni et al., 1981 Tribuzio et al., 1999
G4	Corio and Monastero (Sesia zone)	Gabbronorites	Metapelite and metagranitoids	Lower crust Equilibration condition: 0,6-0,9 GPa-850± 70°C	Undetermined radiometric age, Probably Permian (Chemical affinities)	Exhumation in shallow crustal levels during rifting, suggested by pre-Alpine retrograde metamorphism ((0,25≤P≤0,35 GPa, T<550 °C)	Sesia preserves traces of Jurassic meta-cherts and possibly Triassic carbonates (dolomites)	Rebay and Spalla., 2001 Venturini et al., 1994
G5	Braccia, Fedoz Gabbro (Val Malenco)	Fe-Mg Gabbronorites to quartz diorite	Crust-Mantle transition zone. Intrusion caused granulite metamorphism of Lower crustal (metapelite) and upper mantle rocks.	Lower crust Condition of intrusion: 1-1,2 GPa-1150-1250°C Granulite facies equilibration: 0,9-1,1 GPa, 750°C-850°C	270+6-4 Ma U-Pb on Zircon	Age of pre-Alpine retrograde metamorphism 190-225 Ma $^{40}Ar/^{39}Ar$ on Amphibole, ***Cr*** Exhumation age 170-140 Ma $^{40}Ar/^{39}Ar$ on Amphibole, ***Cr***	Ur Breccia (Jurassic) contains clasts of upper, lower crust and mantle	Hansmann et al., 1996 Müntener et al., 2000 Villa et al., 2000 Trommosdorf et al., 1993
G6	Ivrea zone (Southalpine)	Gabbro-diorite	Granulite facies metapelite	Lower crust Granulite facies equilibration : 0,8-0,9 GPa-950-850 Ma	288±4 Ma U/Pb on Zircon 300-270 Ma Sm-Nd mineral and Rb-Sr on Whole rock methods and mineral isochrons	Pre-Alpine retrograde metamorphism and uplift age 173±4 Ma U-Pb on titanite, ***Cr*** 182±1,6 Ma $^{40}Ar/^{39}Ar$ on muscovite, ***Cr***. (T=380°-400°) 180-230 Ma Rb-Sr and K-Ar on white mica and biotite ***Co***	In Canavese zone, Polymictic breccias (Presumably of early Middle Jurassic age) include clasts of lower crust similar of Ivrea zone	Ferrando et al., 2004 Handy et al., 1999 Henk et al., 1997 Marotta ans Spalla., 2007 Mulch et al., 2002a, b Peressini et al., 2007 Siegesmund et al., 2008 Vavra et al., 1996
G7	External Liguride (Northern Apennine)	Gabbro-norites	No Data, (Sample comes from sedimentary mélange)	Lower crust Granulite facies equilibration : 0,6-0,9GPa-810-920°C	291±9 Ma Sm-Nd plagioclase-clinopyroxene-whole rock	No data	Clastes of Mafic Granulite found in a Santonian-Early Campanian Sedimentary melanges	Marroni and Tribuzio., 1996

As apparent from the brief review above, the late-orogenic early-Permian tholeiitic gabbros widely distributed in lower crustal levels across the entire Alpine domain offer an important tool to constrain the depth of specific crustal sections at the time of intrusion. The formation of these rocks corresponds with Sm-Nd model ages of depleted mantle rocks that were interpreted to represent upper Carboniferous - Lower Permian partial melting residues, predating Mesozoic rifting (e.g. Rampone et al. 1998, Müntener et al. 2004). The granulite to amphibolite facies contact metamorphic aureoles characteristically recorded long-lived isobaric cooling (IBC) following magma emplacement in Permian time, suggesting the attainment of thermal re-equilibration of the crustal section prior to the onset of Jurassic rifting and the onset of exhumation in Jurassic time (Braga et al. 2003; Müntener et al. 2000). These IBC granulites are common in the Alps (Marotta and Spalla 2007) and are interpreted to be the result of major magma underplating (Harley 1989, Bohlen 1991) supporting the idea that the Permian granulites represent true pre-rift lower crust. Importantly IBC granulites should be distinguished from another set of granulitic rocks that are also common in the Alpine basement, which preserve a metamorphic record of isothermal decompression (ITD). Such ITD granulites are interpreted to be derived from thickened crust during the collision and exhumation in a compressional system (Harley, 1989). These granulites (e.g. Sesia zone, Lardeaux and Spalla 1991, Gardien et al. 1994), which have undergone burial and exhumation in an orogenic cycle are consequently not representative of true pre-rift lower crust and may have been in an upper or mid-crustal position already before the onset of rifting (Rudnick and Fountain 1995, Müntener et al. 2000).

4. 2. Mantle

Mantle rocks are, in contrast to continental rocks more difficult to date, and therefore also more difficult to incorporate into complex, time-resolved tectonic scenarios of rifting. Along ocean-continent transitions, it has been recognized over the last 20 years that subcontinental mantle is pulled out underneath the overlying continental crust thereby providing direct access to the mantle and its rift-related history. The best example is found in the Iberia-Newfoundland rift system, where mantle rocks are exposed over tens to hundreds of kilometers at the seafloor (Withmarsh et al. 2001, Müntener and Manatschal 2006). In the Alpine orogenic belt, the mantle rocks are part of ophiolites units, classically interpreted as representing former oceanic lithosphere. Since mantle rocks are found in two different positions in the Alpine tectonic pile, in a supra-Briançonnais (e.g. Liguro-Piemonte ophiolite) and a Subbriançonnais (e.g. possible Valais ophiolites), some authors suggested the existence of two oceanic domains in the Alps (Frisch 1979; Trümpi 1980; Stampfli 1993). However, recent studies question the classical interpretation of a slow-spreading ridge origin for some of the mantle rocks exposed in the Alps (Rampone et al. 1995, Müntener et al. 2004, Manatschal and Müntener 2009). In addition to geochemical and petrological arguments (Rampone et al. 1998, Müntener et al. 2004), the occurrence of mantle rocks underlying crustal rocks, either welded by Permian gabbros (e.g. Malenco; Müntener and Hermann 1996) or sealed by Lower Cretaceous sedi-

Table 1 caption: Compilation of P-T-t data of Permian Gabbros in Alpine/Apennine system ments (e.g. Tasna; Florineth and Froitzheim 1994; Manatschal et al. 2007) indicate that some of these mantle rocks do not originate from accretion at a slow spreading ridge. An alternative interpretation is that some of the Alpine ophiolites may derive from a former Ocean Continent Transition (OCT) (Manatschal and Müntener 2009). Despite the widespread occurrence of MOR-type basalts in Alpine ophiolites, no unambiguous evidence exist for isotopic equilibrium between crustal and mantle rocks, with the possible exception of the Monte Maggiore ophiolite in Corsica (Rampone 2004). In order to emphasize the difference between MOR-type, mature oceanic crust and the Alpine ophiolites, we use the term embryonic oceanic crust, which may best describe the most evolved ophiolites recovered from the Liguro-Piemonte basin (e.g. Chenaillet unit; Manatschal and Müntener 2009).

As shown by Rampone et al. (1998) and Müntener et al. (2004), the mantle rocks within the Alpine Tethys domain have undergone a complex pre-rift history. A key element recorded in mantle rocks is Permian depletion inferred from Sm/Nd model ages on mantle rocks (Rampone et al. 1998), and Sm/Nd mineral isochrones on crustal rocks (Thöni and Jagoutz 1992,1993, Voshage 1988). Thus, some of the mantle rocks within the Alpine Tethys domain probably represent former asthenospheric mantle that underwent fractional melting during Permian time before they cooled and became accreted to the subcontinental mantle lithosphere. The melting event is likely to be related to the Permian magmatic discussed in the previous section (see Fig.4).

During Jurassic rifting, lithospheric extension has caused asthenospheric adiabatic upwelling and decompression melting partially overprinting the pre-rift subcontinental mantle. Müntener et al. (2004; 2009) defined three major types of mantle rocks in the Alpine domain, which can be typified as follow:

An ancient inherited mantle that was at the base of the continental crust before onset of rifting. This is indicated by the fact that mantle rocks are welded to the lower continental crust by a Permian gabbro (e.g. Malenco in Fig. 5d). This type of mantle is made by spinel lherzolites and pyroxenites, equilibrated in the spinel peridotite field in the lithosphere (~800 to 950°C) and is characterized by fertile to depleted compositions. Examples of this type of mantle are Totalp, Malenco, Upper Platta, Tasna (Müntener et al. 2009), part of the External Ligurides (Montanini et al. 2008) and Erro Tobbio (Rampone et al. 2005).

An inherited and infiltrated mantle that is made of plagioclase lherzolites and subordinate harzburgites, equilibrated at high temperature (≥ 1150°C). A Sm-Nd model age on these rocks show that they underwent partial melting related to depletion during Permian time followed by magma infiltration/impregnation and refertilization of this mantle, which is probably of Jurassic age (Rampone et al. 1995, Müntener et al. 2004). The age of this melt impregnation is associated with the development of mantle shear zones that in turn are cut by gabbro dikes dated at 162±2 Ma with U/Pb on zircon (Kaczmarek et al. 2008, Kaczmarek and Müntener 2008). This age relationship indicates that the infiltration post-date the Permian magmatism but predate the emplacement of the Jurassic gabbros.

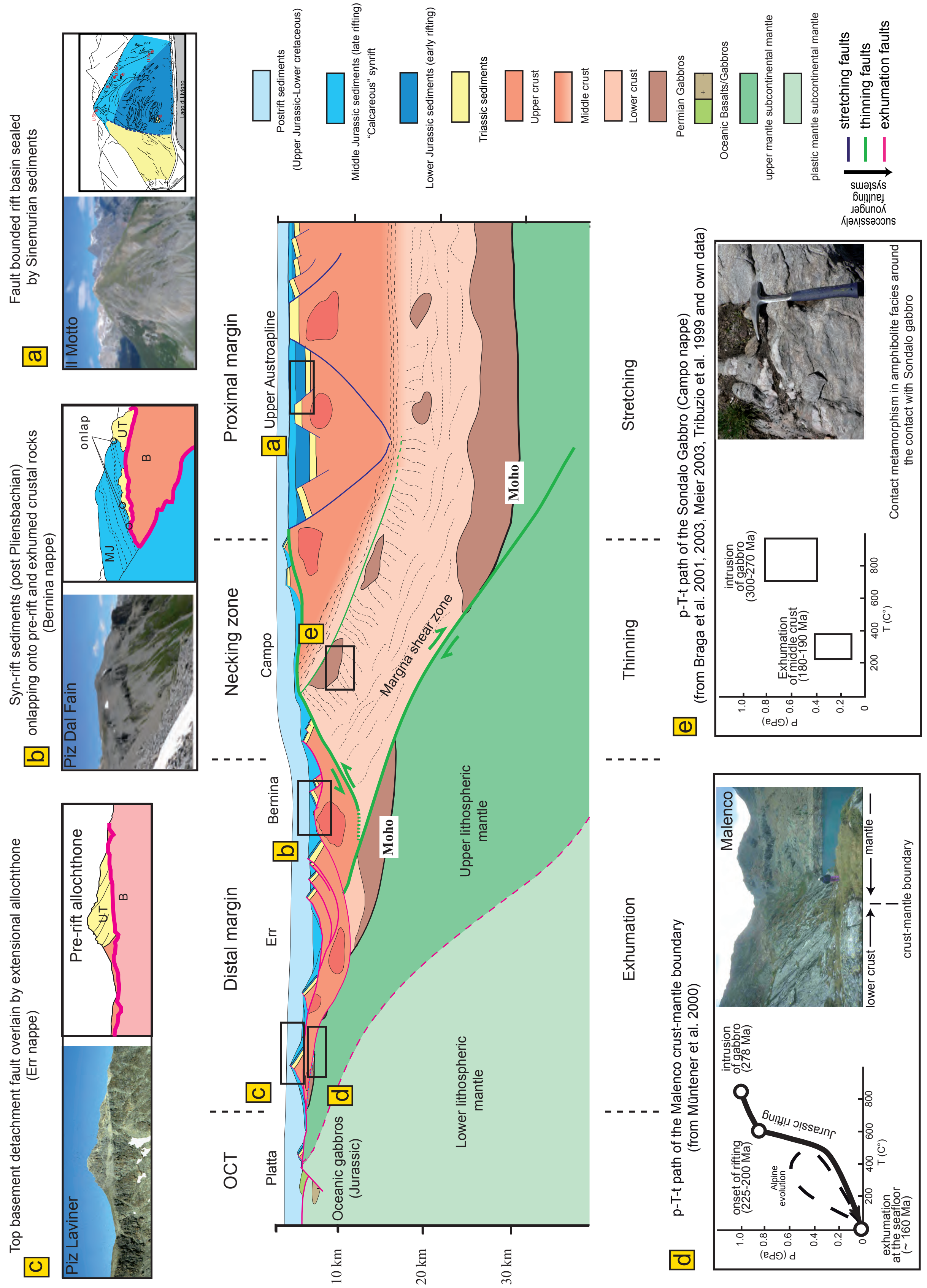
Thus, this impregnation may be related to crustal thinning. Tribuzio et al. (2004) analyzed the Sm-Nd isotopic signature of Mg-gabbros from the Northern Apennines and determined mineral isochron ages between 170 and 179 Ma. Crystallization of Mg-gabbros and mantle refertilization processes are probably closely linked (Müntener and Piccardo 2004). Examples of this type of mantle rocks (e.g. Plagioclase peridotites) are Lanzo, Corsica, Lower Platta, Internal Ligurides, Chenaillet (for references see Müntener et al. 2009).

An oceanic depleted mantle is rare in the Alps and only found in the Monte Maggiore in Corsica (Rampone et al. 2004) and probably in harzburgite-dunite associations in southern Lanzo (Piccardo et al. 2007). This gabbro-mantle association displays similar isotopic relationships as those described from actual mid-ocean ridges (e.g. Salters and Dick 2002). The subdivision into three types of mantle may underscore the complexity of mantle processes during rifting, but it enables to have a first-order criteria to establish the pre-, syn- and post-rift evolution of the mantle during lithospheric extension discussed below.

5. Extensional structures and strain history during rifting

Crustal thinning and extension are accommodated through the activity of several generations of extensional faults and shear zones. Extensional structures differ significantly between proximal and distal parts of the margins. In the classical description of passive rifted margins high-angle faults bounding tilted blocks and limiting half graben type basins are the key structural elements. In the Alps, these structures are widespread and can be found in the proximal European and Adriatic margins. Well-exposed and documented examples for such structures are the Ornon fault bounding the Bourg d'Oisans basin (Chevalier et al. 2003), the Il Motto (Fig. 5c) and the Lugano-Val Grande fault in the proximal Adriatic margin (Bernoulli 1964, Bertotti 1991, Eberli 1988). On a map scale the Lugano-Val Grande fault cuts across the pre-rift sedimentary cover into the basement over a horizontal distance of more than 30 km soling out in a greenschist facies mylonitic shear zone (Bertotti 1991). Thus, these types of structures are decoupled within a quartz-rich middle crust, i.e., at about 10 to 15 km depth assuming normal geothermal gradients of 30°C/km. This is compatible with seismic observations of such structures (e.g. Jeanne d'Arc basin in the Newfoundland margin; Kusznir et al. 1991). In the Alpine Tethys examples, these basins form early in the evolution of the rifted margin and are sealed by Sinemurian sediments (Froitzheim and Eberli 1991, Bertotti et al. 1993) (Fig. 3, 5c).

Although this type of structure remains one of the major building stones of rifted margins, at present it is well accepted that it is not the only structure that can accommodate extensional strain in rifted margins. Low-angle detachment systems can also accommodate a significant amount of extension and eventually lead to mantle exhumation. Early studies suggested that these faults may cut through the entire lithosphere, thus leading to a large scale asymmetry of the margin and to mantle exhumation (Wernicke (1985), Lister et al. (1987), Lemoine et al. (1987) and Boillot et al. (1987). However, well-exposed and documented examples of such detachment fault


c Top basement detachment fault overlain by extensional allochthone (Err nappe)
Piz Laviner
Pre-rift allochthone
UT
B
b Syn-rift sediments (post Pliensbachian) onlapping onto pre-rift and exhumed crustal rocks (Bernina nappe)
Piz Dal Fain
onlap
MJ
UT
B
a Fault bounded rift basin sealed by Sinemurian sediments
Il Motto
OCT
Distal margin
Necking zone
Proximal margin
Platta
Err
Bernina
Campo
Upper Austroalpine
Oceanic gabbros (Jurassic)
Lower lithospheric mantle
Upper lithospheric mantle
Moho
Margna shear zone
10 km
20 km
30 km
Exhumation
Thinning
Stretching
Postrift sediments (Upper Jurassic-Lower cretaceous)
Middle Jurassic sediments (late rifting) "Calcareous" synrift
Lower Jurassic sediments (early rifting)
Triassic sediments
Upper crust
Middle crust
Lower crust
Permian Gabbros
Oceanic Basalts/Gabbros
upper mantle subcontinental mantle
plastic mantle subcontinental mantle
successively younger faulting systems
stretching faults
thinning faults
exhumation faults
d p-T-t path of the Malenco crust-mantle boundary (from Müntener et al. 2000)
intrusion of gabbro (278 Ma)
onset of rifting (225-200 Ma)
Jurassic rifting
Alpine evolution
exhumation at the seafloor (~ 160 Ma)
P (GPa)
T (C°)
Malenco
lower crust
mantle
crust-mantle boundary
e p-T-t path of the Sondalo Gabbro (Campo nappe) (from Braga et al. 2001, 2003, Meier 2003, Tribuzio et al. 1999 and own data)
intrusion of gabbro (300-270 Ma)
Exhumation of middle crust (180-190 Ma)
Contact metamorphism in amphibolite facies around the contact with Sondalo gabbro

systems related to major crustal thinning and mantle exhumation associated with rifted margins are very rare in the geological record. The best examples of exposed rift-related detachment faults are observed in the Eastern Alps (e.g. Err detachment (Fig. 5a; Froitzheim and Eberli 1990; Manatschal and Nievergelt 1997; and Tasna; Florineth and Froitzheim 1994, Manatschal et al. 2007). Indeed these faults differ significantly from those advocated by Wernicke (1985), Lister et al. (1987) and Lemoine et al. (1987), since they do not cut across the entire lithosphere and become active only in a very late stage of rifting. The detachment faults observed in the Alps are characterized by: (1) a top-to-the-ocean transport direction; (2) a fault zone, some tens to some hundreds of meters across, affected by fluid- and reaction-assisted brittle deformation processes and capped by a gouge horizon; (3) depositional contacts with sediments overlying the fault plane at a low angle; and (4) the local occurrence of hanging wall blocks interpreted as extensional allochthons stranded on both exhumed continental and mantle rocks. These faults, referred to as “top-basement detachment faults” (Hölker et al. 2003), are interpreted as exhumed segments of no longer active downward-concave detachment faults which accommodated tens of kilometers of extension and resulted in the exhumation of continental and mantle rocks at the seafloor. Interestingly, the field observations show that these faults were only active in the brittle field, suggesting that they do not penetrate very deep in the crust.

Whitmarsh et al. (2001), Müntener and Hermann (2001), and Manatschal (2004) demonstrated for the Iberia and Alpine Tethys rifted margins that these low-angle detachment faults form in a late stage of the evolution of rifting and post-date major crustal thinning. Lavier and Manatschal (2006) proposed a mode of deformation that they called “thinning mode” which can explain the thinning of the crust to less than 10 km by detachment faults that were decoupled by ductile layers within the crust. These structures are symmetric and decoupled on a margin scale and may thin the crust without producing major fault-related topography. In margins, these deformation structures are expected to occur in the necking zones of the margins where a major change from normal thick crust (± 30 km) to hyper-extended crust (< 10 km) can be observed (e.g. van Avendonk et al. 2009). Examples of such structures have been described in the western Pyrenees (Jammes et al. in 2009). In the Alps, structures that accommodated major crustal thinning and pre-date mantle exhumation are the Pogallo fault in the Ivrea zone (Handy 1987, Hodges and Fountain 1984, Handy and Zingg 1991) and the Margna fault in the Bernina/Margna

Fig. IV. 5: Tentative palinspastic reconstruction of the Adriatic margin across the Austroalpine and South-Penninic units in Grisons (SE Switzerland). The section shows the different type of extensional structures responsable for crustal extension. (a) Il Motto in the proximal margin shows the breakaway of a high-angle normal fault that is sealed by Upper Sinemurian sediments. (b) Piz dal Fain in the necking zone exposes a top basement detachment fault that is overlain by small allochthone of pre-rift sediments onlapped by Late Pliensbachian syn-rift sediments. (c) Piz Laviner shows a tilted block of pre-rift sediments and crystalline basement rocks truncated by a low-angle detachment fault. (d) The P-T-t path of the Malenco crust-mantle boundary shows the exhumation during the rifting of lower crustal rocks close to the seafloor (160 Ma). Note that the gabbros intruded in Permian time in the crust-mantle boundary. (e) Sondalo gabbro was emplaced in a mid-crustal level during the Permian and was exhumed and colled during the transition from early to late rifting (180 Ma).

domain (Fig. 5d) (Hermann and Müntener 1996, Bissig and Hermann 1999, Müntener and Hermann 2001). P-T-t data from the hanging wall and footwall rocks from these shear zones show that they were active during an early stage of rifting and led to significant thinning of the continental crust by at least 15 km (0.4 to 0.5 GPa) (Handy and Zingg 1991; Müntener and Hermann 2001) These majors shears zones were active around 180-200 Ma (182±1,6 Ma 40Ar/39Ar on muscovite for the Pogallo shear zone (Mulch et al. 2002)) at amphibolite to greenschist facies conditions. For the Margna shear zone, the exact ages is less constrained due to the strong Alpine overprint. However the mafic lower crust in the footwall of the Margna shear zone started to be exhumed close to 200 Ma and was at the seafloor close to 160-140 Ma (Müntener et al. 2000, Villa et al. 2000). These shears zones are interpreted to be responsible for the thinning and exhumation of lower crustal rocks during the final rifting. They act as major thinning structure of the continental crust (for more details see table 1 (G5,6)). These fault zones are responsible of the juxtaposition of upper crustal rocks against lower crustal rocks. Despite the good petrological and structural constraints on their activity, it is not yet well understood how these structure accommodated the strain on a lithospheric scale and how they controlled the stratigraphic evolution and distribution of mantle and basement rocks during final rifting; a question that will be discussed below.

6. .Discussion

In the previous sections we reviewed the key features of the sedimentological, structural and crustal and mantle evolution acquired essentially from Permian to Jurassic time in the area that became part of the Alpine Tethys including the Adriatic and European rifted margins. The integration of these different data sets, combined with the knowledge of the different structures that accommodated crustal thinning, provides a detailed picture of the evolution of the Alpine Tethys rifted margins prior, during and after rifting.

The discovery of exhumed continental mantle and hyper-extended crust in present-day magma-poor rifted margins (Boillot et al. 1987; Karner et al. 2007, Contrucci et al. 2004) is at the origin of a paradigm shift within the research of deep water rifted margins. Studies from the Alpine Tethys rift systems show many similarities with the Iberia-Newfoundland rift system (Manatschal and Bernoulli 1999). Indeed these two sets of margins are the only two rift systems from which either drill hole and geophysical data or direct observations enable to observe and describe rift structures and basement rocks from hyper-extended margins. Since more than 20 years, research along both the Alpine Tethys margin and Iberia-Newfoundland rift systems have permitted to observe and confirm: (1) seaward younging of synrift units (e.g. Eberli 1988; Reston 2005); (2) low angle detachment fault showing a top to the ocean sense of shear (e.g. Froitzheim and Eberli 1990; Reston et al. 1996); and (3) mantle unroofing associated with serpentinization (e.g. Pickup et al. 1996, Skelton and Valley 2000, Desmurs et al. 2001). Based on this close link between onshore and offshore studies, new questions emerged and alternative models on the mechanisms leading to lithospheric thinning have been proposed. However, the lack of direct observations from present-day deep-water rifted margins makes it difficult to test these ideas.

This explains the importance of the ancient Alpine Tethys rift system exposed in the Alps, which is the best-studied natural analogue of Atlantic type, magma-poor rifted margins worldwide. In this paper we use this unique data set that was assembled over more than one century to understand the processes of lithospheric thinning and how they are documented in the stratigraphic, petrological and structural record.

Previous studies showed that many rifted margin show some combination of pure and simple shear dominated extension, but it is unclear how, where and when different styles of deformation predominate and how it is documented in the geological record. Although numerical models exist to explain the transition from pure-shear dominated to simple shear dominated systems (e.g. Lavier and Manatschal 2006, Huismans and Beaumont 2008, Regenauer-Lieb et al. 2008) these models need to be rigorously tested and compared with existing, well-constrained data sets. In the following we discuss this transition that is documented in the distal domains of the Alpine Tethys rifted margins occurring from Sinemurian to Bathonian time. We link the subsidence history and sedimentary evolution observed during this phase with the strain distribution and the crustal and mantle processes associated with lithospheric thinning. In a final part, we will discuss how the acquired final architecture in the distal rifted margins may have controlled the subsequent Alpine reactivation.

6. 1. Distal margins in the Alps: stratigraphic record, structural evolution, nature of basement and timing of rifting

Despite the strong Alpine overprint, the record of the Jurassic rifting is still preserved in the rocks and structures within the Alpine orogen. Early studies focused on the external, less overprinted parts of the Alps. These studies enabled to characterize the structure and evolution of the proximal parts of the former rifted margins (Lemoine et al. 1987, Eberli 1988, Bertotti et al. 1993). The restoration of pre-orogenic geometries in more internal parts of the orogen is more difficult. Comparisons with the Iberia-Newfoundland rifted margins provided important constraints and enabled to restore some sections (e.g. Froitzheim and Eberli 1990, Manatschal 2004). These reconstructions showed that the distal margins undergo a polyphase and diachronous evolution with a complex stratigraphic record indicating change in sources, complex subsidence and uplift patterns and a younging of the syn-rift sequences oceanwards. The discovery of syn- and post-rift sediments overlying pre-rift middle and lower crustal and subcontinental mantle rocks in the Alps as well as the reworking of these basement rocks in the sediments (e.g. Tasna Florineth and Froitzheim 1994, Manatschal et al. 2007; Platta Desmurs et al. 2001; Fig. 5) implies the existence of exhumation processes during rifting. The occurrence of low-angle detachment faults that are spectacularly exposed in the most distal margins (Err; Froitzheim and Eberli 1991, Manatschal and Nievergelt 1997; Fig. 5) can explain the exhumation of deeper crustal and mantle levels. The finding of these structures also indicates that during rifting a transition from high- to low-angle faulting had to occur, which is indicated by the younging of the syn-tectonic sediments oceanwards (e.g. Fig. 3 and 5). The change in the deformation style results in the transition from a crustal architecture characterized by tilted blocks in the proximal margin to the extensional allochthons in the distal margin, which is well recorded on a transect across the Adriatic margin (Fig.

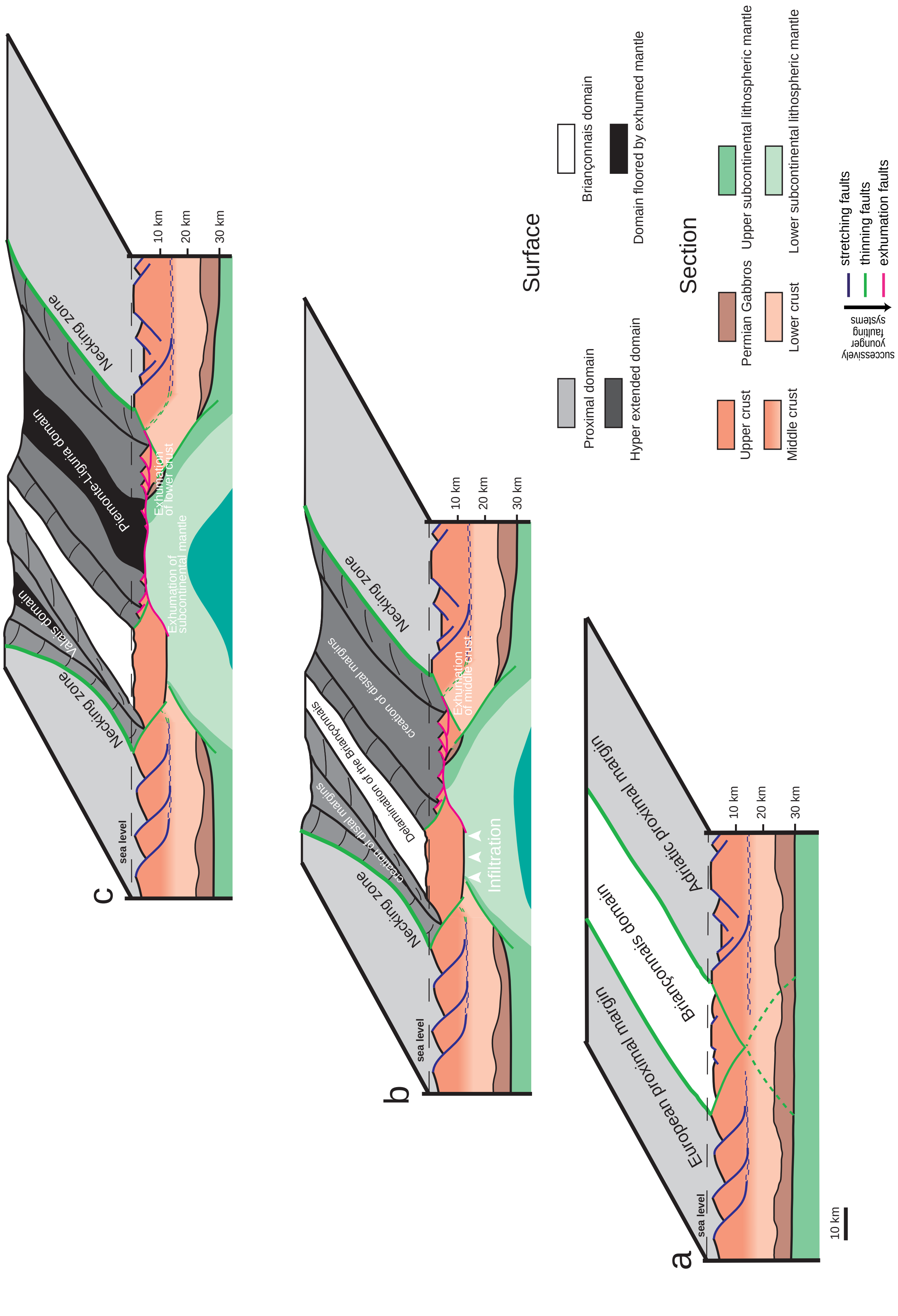
5). Furthermore, exhumation processes resulted in the formation of new paleogeographic domains floored by exhumed crustal and mantle rocks. Although in the Alps it is difficult to estimate the total width of the area floored by newly exhumed basement rocks, in Iberia this domain is up to 200 km wide (Péron-Pinvidic and Manatschal 2008). Thus, exhumation processes in distal margins forms paleo-geographic domains that did not exist prior to rifting (Fig. 6). In the Alps, these are the Proto Valais and Piemonte-Liguria domains, which are characterized by structures, basement rocks and sedimentary sources and architecture that are different from those observed in the adjacent proximal European and Adriatic rifted margins. The interpretation that these domains may have essentially evolved during final rifting, as shown in Figures 6 and 7, has major implications for the interpretation of the rift and post-rift history and the subsequent Alpine reactivation.

Classically, rifting is dated by using stratigraphic methods dating the syn-rift sequence and the first sediments overlying oceanic and crustal units and crystallization ages of syn-rift magmatic rocks, which are in the case of the Alps MOR-type gabbros emplaced in the OCT. Using this approach, rifting within the Proto Piemonte-Liguria domain initiated in latest Triassic to early Jurassic time and earliest mantle exhumation occurred in Bathonian to Callovian time, as indicated by the convergence of magmatic and stratigraphic ages (Bill et al. 2001, Manatschal and Müntener 2009). The dating of rifting and first mantle exhumation in the Valais domain is much less constrained (due to the Alpine overprint) and was mainly based on the observation that Lower Cretaceous sediments are observed to seal exhumed mantle in the Tasna OCT (Florineth and Froitzheim, 1994). However, the sealing of exhumed basement by sediments only provides a minimum age and does not necessarily date the exhumation as such. In this context it is important to note that in the Piemonte-Liguria domain contacts between Lower Cretaceous deep water sediments and exhumed mantle and crustal rocks can be found in the Platta (Desmurs et al. 2001) and Bernina nappes. This is in line with observations made along the Iberia margin were post rift sediments onlap onto exhumed basement highs resulting in a hiatus of up to 60 Ma between the age of basement exhumation (dated with 40Ar/39Ar method (Jagoutz et al. 2007) and the first preserved sediments (dated by microfossils) (see ODP Sites 900, 1067 and 1068 in Iberia; Wilson et al. 2001). Thus, the Lower Cretaceous sediments on top of exhumed mantle in the Tasna nappe cannot be used to date the age of mantle exhumation (e.g. Florineth and Froitzheim 1994). Although the ages of the ophiolites and sediments of the Proto-Valais domain (see above) are not as well determined as for the Piedmont-Liguria domain, the available data is more compatible with a late Middle to Late Jurassic evolution of this basin (see discussion in Manatschal and Müntener 2009), rather than an Early Cretaceous opening, for which there are neither stratigraphic nor isotopic constraints, expect for minor mafic magmatism locally preserved in the Central (Liati et al. 2003, 2005) and Western Alps (Liati and Froitzheim 2006).

The discovery of major exhumation surfaces in the Proto-Valais and Piemonte-Liguria domains (e.g. Tasna, Err, Platta; Manatschal 2004) makes that rifting in these domains can be dated also by looking at the age of first sealing of exhumed basement and cooling ages of exhumed middle and lower crustal rocks. 40Ar/39Ar cooling ages obtained from lower and middle crustal rocks from the Ivrea zone (180-210 Ma) (e.g. Handy et al. 1999 and references therein), Malenco crust-mantle boundary (exhumation start from 200 Ma and exhumation at seafloor are between 170-140 Ma) (Müntener et al. 2000, Villa et al. 2000) and Campo (180-200 Ma) (e.g. Meier 2003) show a major cooling/deformation event at around 180-200 Ma (for references and more details see table 1). The overlap between stratigraphic and cooling ages suggests that the transition between distributed stretching and localized thinning and exhumation initiated during late Sinemurian time. At this stage, rifting in the proximal margins stopped, the basins were sealed (e.g. Il Moto Eberli 1988; Fig. 5), Bourg d'Oisans, Chevalier et al. 2003; Southern Alps, Bertotti et al. 1993) and deformation localized in the more distal domains (e.g. Fig. 6a and b).

Contemporaneous extensional deformation occurring at amphibolite to greenschist facies conditions in the middle crust has been reported from the Pogallo shear zone and the Margma faults, which accommodated significant crustal thinning (Handy et al. 1999, Müntener and Hermann 2001). The major activity of the Pogallo shear zone, which is found in the Southern Alps, has been constrained at ca. 182.0±1.6 with 40Ar/39Ar geochronology on white mica (Mulch et al. 2002) A key observation in the Alpine Tethys domain is the uplift and sub-aerial exposure of the Briançonnais domain during Jurassic rifting. This emersion has been dated as Toarcian-Aalenian to Bathonian-Oxfordian based on sedimentological and paleontological arguments (Faure and Mégard-Galli 1988 and references therein)

Finally, the Sm-Nd mineral isochrones in plagioclase peridotite and Mg-gabbros within exhumed mantle rocks from Liguria and the northern Apennines observed in the Liguria-Piemonte domain were dated by Rampone et al. (1995) and Tribuzio et al. (2004) to 164±10 Ma and 180-173 Ma, respectively. These ages are presumably related to melt infiltration and cooling prior to mantle exhumation (see Müntener and Piccardo 2004). The infiltrated mantle rocks are intruded by gabbros (Lanzo: 161±2 to 158±2 Ma; Kaczmarek et al. 2008; Platta: 161±1 Ma, Schaltegger et al. 2002) demonstrating that melt infiltration pre-dates their emplacement. On the other hand, the observation that the infiltrated rocks are overprinting previously depleted mantle rocks (depletion dated as Permian; Rampone et al. 1998, Müntener et al. 2004 or Proterozoic; Bodinier et al. 1991), support the interpretation of Müntener and Piccardo (2004) that infiltration postdates Permian magmatism and pre-dates mantle exhumation in late Middle Jurassic time.


a
b
c
sea level
10 km
20 km
30 km
European proximal margin
Briançonnais domain
Adriatic proximal margin
Necking zone
Infiltration
Exhumation of middle crust
creation of distal margins
Delamination of the Briançonnais
Valais domain
Piemonte-Liguria domain
Exhumation of lower crust
Exhumation of subcontinental mantle
Surface
Proximal domain
Hyper extended domain
Briançonnais domain
Domain floored by exhumed mantle
Section
Upper crust
Middle crust
Permian Gabbros
Lower crust
Upper subcontinental lithospheric mantle
Lower subcontinental lithospheric mantle
successively younger faulting systems
stretching faults
thinning faults
exhumation faults

In conclusion, we start to see a convergence in the timing of onset of: (1) the localization of rifting in the distal margin, (2) the change in the style of deformation from pure to simple shear leading to exhumation of mid-to lower curstal rocks, (3) the infiltration of the underlying lithospheric mantle, and (4) the uplift of the Briançonnais domain. We assume that these events were interconnected and responded to a major re-organization of rifting that occurred in late Sinemurian time. At the surface, this event resulted in the creation of new paleogeographic domains, the Proto Valais and Piemonte Liguria domains, and at depth with the onset of the rise of the asthenosphere (Fig. 6).

6. 2. What happens in distal margins prior to mantle exhumation?

As shown and discussed before, rifting in the Alps is poly-phase (Fig. 6) and a major change in the mode of extension from high- to low-angle dominated fault systems began in late Sinemurian time. This is reflected by (1) the migration of the deformation into the future distal margin, (2) a change in the sedimentary system from carbonate dominated to mainly siliciclastic syn-rift sequences, (3) the exhumation of middle and lower crustal rocks in the footwall of major extensional faults, (4) the impregnation of mantle rocks, and (5) the uplift of the Briançonnais domain that is followed by rapid subsidence (e.g. Lemoine et al. 1986; Borel 1995). All these processes post-date initial rifting (stretching) and predate mantle exhumation, i.e. they occur while the lithosphere is undergoing extreme thinning. During crustal thinning, it appears that extension is localized at the edges of a paleogeographic domain, the Briançonnais domain, which preserves the pre-rift stratigraphic cover and, interestingly, is uplifted during extension (Fig. 6). In between this domain and the future Adriatic and European margins, two new paleogeographic domains develop, referred to as the Proto Piemonte Liguria and Proto Valais domains (Fig. 6). The stratigraphic record shows a strong subsidence from the Late Pliensbachian onwards in these two domains that contrast with the uplift of the Briançonnais domain (Fig. 3). Continentwards, these domains are limited by areas that are exhumed, uplifted and eroded (e.g. Bernina, Campo, Grosina) on the Adriatic margin (Fig. 5); and probably also the most internal parts of the external massifs (Argentera, Pelvoux) at the transition between the proximal and distal European margin (Fig. 1). The creation of such new paleogeographic domains is interpreted to coincide with the formation of the necking zones of the future margins between the Briançonnais domain and the two future margins, and lead to the formation of two domains in which detachment systems are thinning the crust. Thus, these two domains are formed by exhumed basement that is overlain by remnants of upper crust and pre-rift sediments (e.g. allochthons) derived from the delamination of the Briançonnais domain (e.g. Fig. 6).

Fig. IV. 6: Conceptual 3D model showing the temporal and spatial evolution of the Alpine domain from onset of rifting to final rifting. (a) Early rifting (stretching) phase characterized by distributed fault bounded basins that can be found across the whole future margin. (b) Late rifting (thinning) phase during which the extension is localized in the Proto-Valais and Piedmont-Liguria domains. Note that during this stage the distal margin is undergoing extreme crustal thinning. Note also that the Briançonnais domain is delaminated and uplifted during this stage. (c) Onset of mantle exhumation in the Valais and Piedmont-Liguria Basins. Note that the Valais is interpreted to die out laterally, whereas the Piedmont-Liguria basin is evolving into an embryonic oceanic domain.

This evolution, as discussed above, can explain most of the observed stratigraphic contacts and the change in the sedimentary sources (Fig. 3), the occurrence of top basement detachment faults and extensional allochthons, as well as of mid and lower crustal rocks and mantle rocks in direct contact to sediments (Fig. 5). The uplift of the Briançonnais domain and its subsequent rapid subsidence may be related to the rise of hot, infiltrated lithospheric mantle underneath the extended domain (e.g. Müntener et al. in press). The rise hot and less viscous mantle rocks underneath extending lithosphere would not only explain the isostatic evolution of the Briançonnais domain, but also strain localization and occurrence of infiltrated mantle within the OCT. During this event, middle and lower crustal rocks are exhumed at the seafloor and overlain by extensional allochthons derived from the delamination of the Briançonnais domain. These processes may result in a final architecture characterized by mantle windows surrounded by extensional allochthons (e.g. Fig. 9 in Péron-Pinvidic et al. 2007) near the continent, and further oceanwards, by exposure of continuous mantle domains forming the floor of the OCT. From our observations, this process occurred on both sides of the Briançonnais domain; however, it appears that extension in the Valais domain (Steinmann 1994) stopped earlier while mantle exhumation in the Piemonte Liguria domain continued into an embryonic oceanic stage. From available data (see compilation in Manatschal and Müntener 2009 and Beltrando et al. 2010) these processes occurred in late middle to early Late Jurassic time, which is also supported by the stratigraphic record within the two basins.

6. 3. Implication for paleogeographic interpretations

The re-interpretation of the rift evolution in the distal margins has important consequences for the paleogeography of the pre-rift position of distal domains and the extent and nature of oceanic domains. Within areas undergoing extreme crustal thinning, there is increasing evidence that top-basement detachment faults can exhume crustal and mantle rocks over tens to hundreds of kilometers at the seafloor (e.g. ODP drilling results along the Iberia and Newfoundland rifted margins, Péron-Pinvidic and Manatschal 2008). Although these domains did not exist before rifting, pre-rift sediments may exist within these basins as extensional allochthons. However these units are not continuous and they overlie exhumed crust or mantle. This observation has major implications not only for the paleogeographic reconstructions, but also for the structural interpretation of such units in collisional orogens. As an example, in many Alpine reconstructions, the lack of continuity between Triassic units is interpreted as the result of Alpine deformation. Alpine geologists used Triassic units to restore the former pre-Alpine top upper crust however in the most distal part of the margin the Triassic pre-rift cover was already discontinued during the rifting due to the activity of low-angle detachment faults.

Rift domains in present-day hyper-extended rifted margins show a complex, polyphase evolution, which results in a 3-D rift architecture resulting in V-shaped basins, continental ribbons, H-blocks, extensional allochthons, exhumed mantle domains and oceanic crust (e.g. Péron-Pinvidic and Manatschal 2009). While these units/structures are defined and can be recognized in present-day margins it is more difficult to recognize them in orogens. V-shaped basins are located on both sides of the southern North Atlantic (e.g. Rockall Trough, Porcupine, Orphan or Bay of Biscay; e.g. Péron-Pinvidic and Manatschal, 2009). They have similar sizes like the Alpine Tethys basin, and are locally floored by mantle rocks and even in some case by extrusive magmatic rocks. H-Blocks (see Lavier and Manatschal 2006, Péron-Pinvidic and Manatschal 2009) are observed within these basins and correspond to thinned crust that is separated from the surrounding continents (proximal margins) by major conjugate thinning structures (necking zones). In V-shaped basins were the pre-breakup configuration is preserved like in the southern Orphan basin; (Péron-Pinvidic and Manatschal 2009), the H-Block lies between the two continents, similar to the situation shown in Fig. 6a where the Briançonnais lies between Europe and Adria. Further to the north in the Orphan basin, were the basin was more extended, it can be observed that the H-block is thinned at its margins and exhumation structures develop, leading to the exhumation of lower crustal and mantle rocks flooring deeper basins (Péron-Pinvidic and Manatschal 2010). This situation is very similar to the situation shown in Figures 6b and 6c, where the Proto Valais and Piemonte-Liguria domains develop between the Briançonnais (equivalent to the H-Block in Orphan) and the adjacent European and Adriatic proximal rifted margins. Thus, we interpret the Valais, Briançonnais and Piemonte-Liguria domains, in analogy with observations made on present-day rifted margins as the result of a complex rift history that developed essentially between late Sinemurian and Callovian time between the proximal European and Adriatic margins.

6. 4. Structural, compositional and thermal rift inheritance: implications for the Alpine reactivation

As shown in Figures 6c and 7c and discussed in the previous section, the paleogeography at the end of rifting was complex and strongly 3D. Although the width of the newly created domain is difficult to estimate, we suggest that its total width was in the order of 800 kilometers across a section shown in the northern part of Fig. 6c, which is a rough approximation assuming about 45myr of extension/accretion at < 2cm (for more detailed discussion see Manatschal 2004). It is, however, evident that the amount of extension cannot be determined very accurately and cannot be extrapolated across the whole domain.

It seems obvious that a complex palaeogeography as shown in Fig. 6c exerted a first-order control on the compressional evolution of the orogen, although this latter statement is difficult to proof. The major problem is that paleogeographic reconstructions are dependent on the interpretation of orogenic structures which is particularly true for reconstructions that result from the kinematic inversion of the orogenic structures (e.g. Schmid et al. 1996) or a combination of petrological and geochronological data (e.g. Berger and Bousquet 2008, Froitzheim et

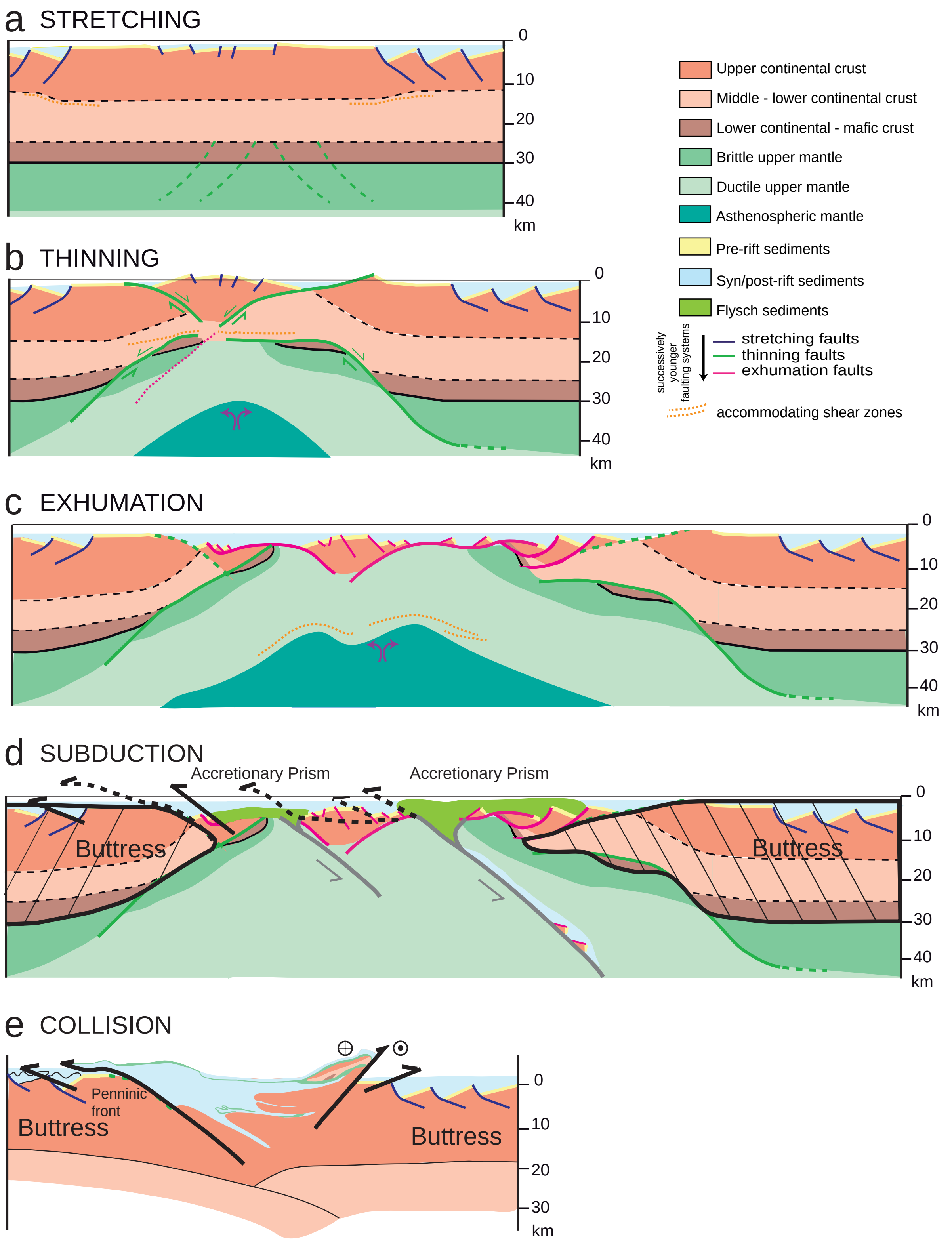


Fig. IV. 7: Conceptual 2D model showing the evolution from (a) an early rifting (stretching) phase, to (b) a late rifting (thinning) phase, to (c) an mantle exhumation phase, to (d) onset of subduction and (e) final collision. Note that two subduction zones form in the two basins within the two OCT zones. Note also that the necking zones become the buttress during final collision and that the thinning faults are reactivated during collision (for further discussion see text).

al. 1996). Moreover, the accessible units preserved in collisional orogens may provide a highly biased view of the pre-compressional structures, since a major part of the geological record related to rifting and seafloor spreading may have been either reactivated or subducted. Nevertheless, scientific results from the last 20 years comparing present-day and fossil rifted margins (e.g. Boillot and Froitzheim 2001, Manatschal 2004) are encouraging in that the structures that we can find and interpret in figures 3 and 5 are comparable to those seismically imaged and drilled off Iberia and Newfoundland. Therefore, we infer that the paleogeographic reconstruction shown in Figure 6c may represent a likely scenario for the pre-collisional situation in the Alpine domain.

Although we are aware about the difficulty of details about how the rifted margins were reactivated during Alpine compression, we outline two points that merit a more thorough discussion: 1) do the necking zones act as buttress during continental collision and 2) where did the subduction zones initiate, and what is the evidence for an oceanic slab in the Alps. A first important observation is that the major change in the style of Alpine deformation coincides with the transition between the former proximal and distal rifted margins. In the Alps, this corresponds to the limit between internal and external units and corresponds to the location of the Penninic Front. This structure represents a complex polyphase structure (Ceriani et al. 2001) separating the Penninic from the European units (fig 7d, e). On deep seismic sections, this structure can be followed into the mantle (Schmid et al. 1996), which highlights the importance of this structure. The facies of the Upper Jurassic and Cretaceous sediments in the hanging and footwall of this fault are different, juxtaposing distal units over proximal units. In the present interpretation, this structure is interpreted as a major Alpine thrust fault, however, it does not explain why this structure can be followed into the mantle. An alternative interpretation would be that this structure represents a reactivated rift structure that was responsible for the thinning of the crust within the necking zone of the former margin (e.g. Fig. 7d). Such an interpretation could explain the mantle exhumation observed in some parts of the Valais basin as well as the Alpine structures. During final compression this fault was reactivated and the necking zone of the European margin acted as a buttress (Fig. 7d).

Although the existence of a subduction zone in the Alps seems to be supported by geophysical and petrological data, the nature and size of the slab as well as the processes controlling onset of subduction are still little understood. The classical models assume the existence of oceanic crust in the Alps and assume that the subduction initiated in the southeastern margin of the ocean. However, it is important to mention that at present no unambiguous oceanic crust was found in the Alps. All mantle rocks are subcontinental and the age of the MOR gabbros are older than 140 Ma (Manatschal and Müntener 2008). Thus, two hypotheses exist for the oceanic domains in the Alps. Either that the oceanic crust existed and was subducted, or that it never formed, which would explain the lack of evidence for unambiguous oceanic crust. In present-day rifted margins OCT are in the order of 200 to 300 km wide. It is also important to mention, that basins like Orphan, Rockall Trough or Porcupine in the southern North Atlantic show wide domains of exhumed mantle and embryonic crust. Thus, also at present we cannot answer to the question if we deal with an exhumed mantle domain or a classical oceanic lithosphere in the Alpine domain, it may be important to think and compare the Alpine reconstructions with marginal, V-shaped basins rather than with big Atlantic oceans. V-shaped basins may explain better the complex, small-scale variability of the system, which is, to such an extend not observed in present-day oceans.

7. Conclusion

In this paper we reviewed the key features of the sedimentological, crustal and mantle evolution acquired essentially from Permian to Jurassic time in the area that became part of the Alpine Tethys including the Adriatic and European rifted margins. The integration of these different data sets, combined with the knowledge of the different structures that accommodated crustal thinning, provides a detailed picture of the evolution of these rifted margins prior, during and after rifting and enables to address three major questions: (1) how and when did the lithosphere thin; (2) what are the structures that document this lithospheric thinning and (3) how are these processes recorded in the structural, stratigraphic and petrographic record.

We can demonstate that crustal thinning occurred by localized large-scale fault and shear zones that were thinning the lithosphere and exhumed lower crustal and mantle rocks at the seafloor. Using stratigraphic and isotopic ages we can demonstrate that these major structures initiated in late Sinemurian when the proximal basins were sealed and deformation migrated in the distal parts of the future margin, probably simultaneous with infiltration of MOR magmas in the lithospheric mantle and the uplift of the Briançonnais domain. As a consequence of the change in style of deformation from pure shear to simple shear, the extensional structures differ significantly between proximal and distal parts of the margins. While in the proximal margins classical high-angle fault bounded tilted blocks with half graben type basins are the predominant structure, in the distal margins top-basement detachment faults with extensional allochthons can be observed. Exhumation resulted in the formation of new paleogeographic domains, the Proto Valais and Piemonte Liguria domains, floored by exhumed crustal and mantle rocks and locally also by embryonic oceanic crust.

The re-interpretation of the rift evolution in the distal margins in the Alps bear important insights in the evolution of rifted margins in general and has also major implications for the paleogeographic interpretation of the Alpine domain and the subsequent Alpine compressional overprint.

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DISCUSSION

Mon travail de thèse a eu pour but de caractériser et comprendre les mécanismes d'extension lithosphérique qui vont former les marges passives peu magmatiques. Ces mécanismes montrent le passage d'une croûte continentale plus ou moins équilibrée (~30 km) dans les parties proximales, à une croûte très amincie dans les parties distales (~10 km) de la marge. Il est établi que dans les marges passives peu magmatiques la transition entre les deux domaines s'effectue dans la zone d'étranglement appelée « necking zone », cependant les données acquises sur les marges actuelles sont insuffisantes pour caractériser ces mécanismes dans l'espace et dans le temps. Ainsi, mon étude a été menée sur les reliques de la marge Adriatique de la Téthys Alpine, préservée dans les Alpes.

Cette thèse est ainsi focalisée sur les domaines des nappes Austroalpines et Sud-Pénnique, dans le Sud-Est de la Suisse et le Nord de l'Italie. Au sein de ces nappes, les reliques de la marge proximale ont été identifiées dans la nappe Austroalpine supérieure (Froitzheim 1988; Conti et al. 1994) tandis que les nappes Austroalpines inférieures et Sud-Pénnique représentent respectivement le domaine distal et la transition océan-continent (Froitzheim et Eberli, 1990; Manatschal et Nievergelt 1997 ; Desmurs et al. 2001).

Les unités de Bernina-Campo-Grosina sont situées à l'interface entre ces deux domaines distincts qui n'avaient été que peu étudiés. Les études entreprises dans cette zone ont permis la description et l'identification des structures Alpines, la reconstruction de la position paléogéographique de ces unités dans la marge passive et l'identification des structures pouvant accommoder cet extrême amincissement crustal au cours du rifting.

La caractérisation de la géométrie Alpine a permis la rétrodéformation de ces unités pour en fixer les positions dans la marge Adriatique. Ainsi l'unité de Bernina possède les reliques d'une marge distale, tandis que les unités de Campo-Grosina représentent une ancienne « necking zone ».

La terminologie utilisée dans ce travail, ainsi que la chronologie du rifting dans le domaine de la Téthys Alpine sont décrites dans la figure 1.

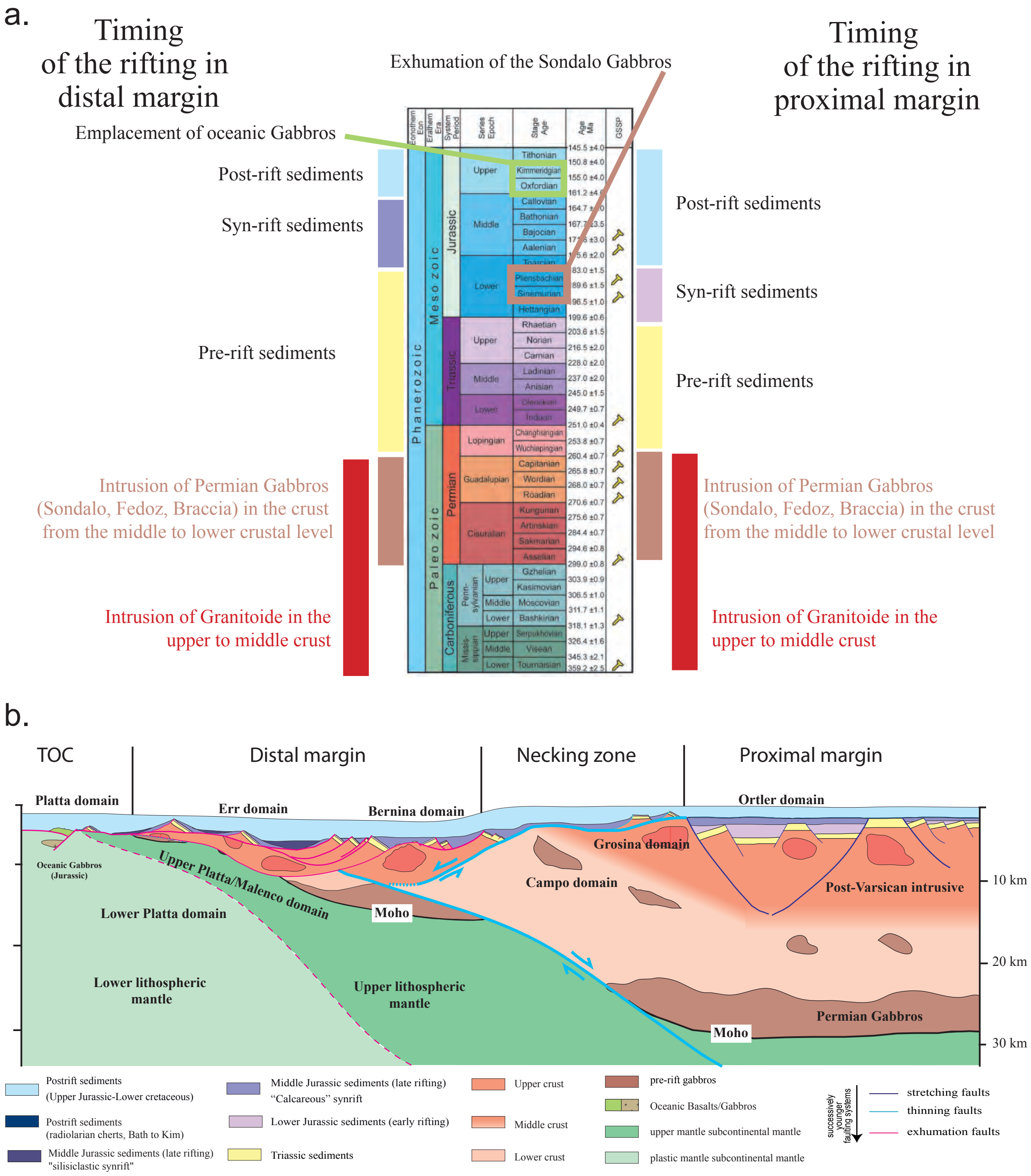


Fig. V. 1: (a) Charte stratigraphique internationale modifiée d'après Gradstein et al. 2004, montrant les principaux événements tectoniques affectant les domaines proximaux et distaux de la marge Adriatique de la Téthys Alpine. (b) Reconstruction de la marge Adriatique au Jurassique supérieur montrant la terminologie et la position des principales unités utilisées dans cette thèse.

1. Répartition spatiale des différents niveaux de croûte dans la zone d'étranglement (« necking zone ») ainsi que dans les parties proximale et distale de la marge *(Publication 2)*

Les marges fossiles Téthysiennes permettent d'avoir accès aux roches formant les domaines proximaux, distaux et les « necking zones » des marges passives peu magmatiques. Il est ainsi possible de déterminer la lithologie, et éventuellement la position, des roches composant les différents domaines au cours du temps, c'est à dire avant et pendant le rifting. La détermination des différentes unités lithologiques présentes avant le rifting est rendue possible par la marque de l'événement thermique du Permien ayant entrainé l'intrusion de corps acides et mafiques dans toute la croûte (Schaltegger et Brack 2007 ; Schuster et Stüwe 2008). Les études pétrologiques sur ces intrusions et leurs auréoles métamorphiques permettent d'établir leur profondeur de mise en place dans la croûte au Permien. De cet événement Permien jusqu'à la fin du Trias, aucune phase tectonique majeure susceptible d'exhumer des portions de croûte continentale dans cette zone n'a été enregistrée. Dans le domaine Alpin, l'enregistrement sédimentaire au cours de cette période et jusqu'à la fin du Trias se traduit par l'établissement de deux plateformes carbonatées successives et sans indice d'activité tectonique majeure (Lemoine et al. 1986 ; Bertotti et al. 1993). Ces observations sont complétées par une étude pétrologique sur une relique de croûte inférieure permienne ayant subi une évolution isobare jusqu'à la fin du Trias (Müntener et al. 2000), au début du rifting.

Les reliques de la marge Adriatique, dans les nappes Austroalpines et dans le Sud des Alpes, m'ont permis de caractériser l'architecture crustale ainsi que l'évolution des différents niveaux de la croûte, au cours du rifting. Cependant, par l'incorporation de la marge dans l'orogène Alpine, seule une partie de cette marge passive a été préservée. Ainsi l'échantillonnage de ces domaines préservés de marge, parfois partiel est complété par les données géophysiques obtenues sur les marges passives actuelles qui permettent de caractériser leur architecture et évolution crustale à grande échelle.

1. 1. Les nappes Austroalpines supérieures (unité d'Ortler): reliques des zones proximales de marge

Dans le domaine Téthysien Alpin, les reliques des parties proximales montrent des structures extensives impliquant principalement la partie supérieure de la croûte (e.g. Froitzheim 1988 ; Bertotti et al. 1991, 1993). La couverture pré-rift repose sur le socle paléozoïque en contact primaire. Dans ces domaines il n'est cependant pas possible de déterminer l'épaisseur crustale ni d'avoir accès à l'architecture des zones les plus profondes. Ces lacunes peuvent être comblées par les données acquises sur des marges actuelles. Les études de sismique réfraction et réflexion montrent que, dans des domaines similaires, la croûte n’a été que faiblement ou non amincie, indiquant ainsi une croûte équilibrée d'épaisseur de l'ordre de 30 à 35 km et associée à un Moho relativement plat (Lau et al. 2006).

1. 2. Les unités de Campo-Grosina : relique d’une « necking zone »

De nombreuses marges actuelles présentent, dans la direction de l’océan, une zone de près de 50 à 60 km de large montrant un amincissement brutal de la croûte qui diminue de 30 km à 10 km. Ceci se traduit par une forte remontée du Moho, marquée par un angle pouvant atteindre 35° (Lau et al. 2006). Cependant peu d’informations supplémentaires peuvent être apportées par l'étude des marges actuelles, du fait du peu de données géophysiques disponibles ainsi que de l’absence de forage profond.

Les unités de Campo-Grosina sont situées entre les reliques des zones de marge proximales et distales et sont interprétées comme corrrespondant à une « necking zone ». Ces unités représentent la section crustale de 3 km au niveau d'une « necking zone » établie au Jurassique Elles peuvent donc donner des informations sur la structure de ce domaine qui accommode l’amincissement lithosphérique au cours du rifting.

L’unité de Campo est composée d’un socle paléozoïque essentiellement constitué de métapélites équilibrées dans le faciès amphibolite supérieur et affectés par une forte foliation verticale avant le Permien. En effet, au Permien le socle de Campo a été intrudé par de nombreux corps mafiques et acides. L’un de ces corps est le gabbro de Sondalo qui a fait l’objet de nombreux travaux (Tribuzio et al. 1999 ; Braga et al. 2001, 2003). Il a été démontré qu'il se met en place entre 270 et 300 Ma (Tribuzio et al., 1999) dans un niveau intermédiaire de la croûte, à une pression de 0,8 GPa et une température estimée à ~900°C (Braga et al. 2003) (Fig. 2a). Ces données indiquent que l’unité de Campo était à une profondeur de 20 à 30 km au Permien. Müntener et al. (2000) ont étudié l’évolution de la transition entre croûte continentale et manteau subcontinental scellé par un gabbro Permien (à des pressions de 1-1,2 GPa et des températures de 1150-1250°C) dans l’unité adjacente de Margna. Il a été possible de démontrer que, pendant le Permo-Trias avant le début du rifting, cette portion de croûte inférieure a subi une évolution isobare caractérisée par une pression de 0,8±0,1 Gpa et une température de 600±50°C (Fig. 2c) (Müntener et al. 2000). Ainsi, il est proposé que l’événement thermique Permien et le rifting Jurassique sont deux phases indépendantes. Il est donc raisonnable de considérer que le gabbro de Sondalo n’a pas subi de mouvement tectonique notable depuis sa mise en place au Permien jusqu’au début du rifting. L’unité de Campo comprenant le gabbro de Sondalo était donc placée à un niveau intermédiaire de la croûte, avant le rifting. Les datations des roches encaissantes proches de l'intrusion de gabbro par la méthode $^{40}Ar/^{39}Ar$ sur biotite, entreprises durant cette thèse, montrent un âge de refroidissement de 184,55±1,22 Ma et 188,67±0,86 Ma. Ces données démontrent l'exhumation de cette croûte moyenne lors du rifting au début du Jurassique (~180 Ma). Il est ainsi établi que l'exhumation de niveaux crustaux plus profonds s'effectue au sein de la « necking zone » pendant le rifting.

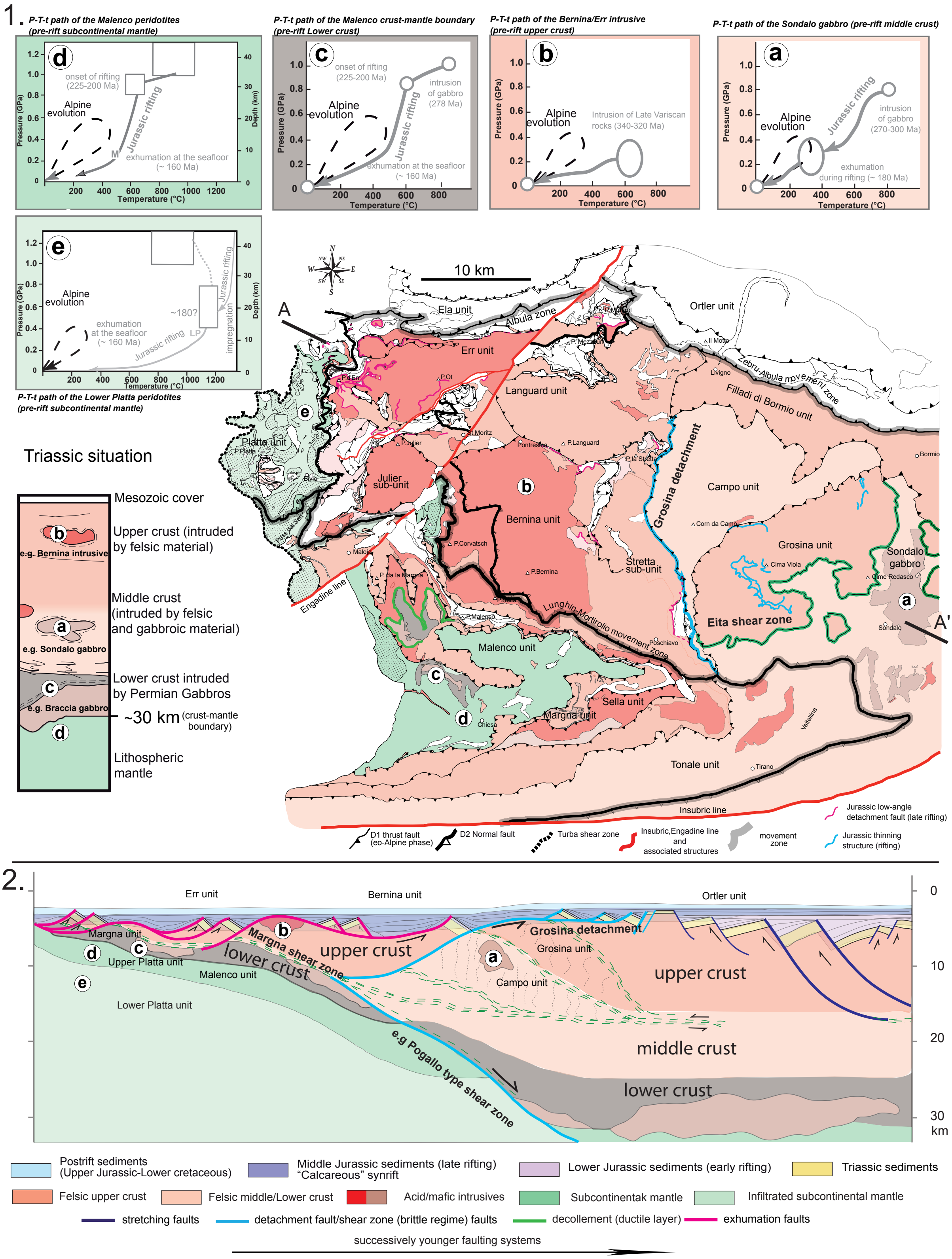
J'ai aussi établi que l'unité de Grosina qui surplombe l'unité de Campo est constituée majoritairement de croûte supérieure, dès le rifting. Cependant aucune couverture sédimentaire mésozoïque n'est préservée dans cette unité. Mais l'unité a été intrudée par des granodiorites certainement d'âge Permien, comparable à celui des intrusions décrites dans l'unité voisine de Bernina (Von Quadt et al. 1994) qui se sont mises en place dans les niveaux superficiels de la croûte.

Ainsi l'ensemble de ces nouvelles données ont rendu possible la reconnaissance, la caractérisation des lithologies présentes dans une « necking zone » fossile.

1. 3. Les nappes Austroalpines inférieures (unité d'Err-Bernina-Margna) : reliques d'une marge distale

L'unité de Bernina ainsi que les unités adjacentes (unité de l'Err) ont été intrudées à la fin de l'orogène Varisque jusqu'au Permien par des roches acides associées à un magmatisme extrusif caractérisé par la présence de rhyolite (Von Quadt et al. 1994). Ce magmatisme extrusif démontre que l'unité de Bernina représentait une croûte supérieure dès le Permien (Fig. 2a).

Dans l'unité adjacente de Margna, précédemment considérée comme une relique de marge distale (Hermann et Müntener 1996 ; Hermann et al. 1997), il a été possible de montrer la présence de croûte inférieure pré-rift scellée au manteau subcontinental par un gabbro permien. Au sein de cette unité, une section crustale, bien que déformée lors de l'orogènèse Alpine, montre la juxtaposition de croûte inférieure en contact franc avec la croûte supérieure le long d'une zone de cisaillement majeure (la zone de cisaillement de Margna (Hermann et Müntener, Bissig et Hermann 1997)) (Fig 2) et donc avec une lacune des niveaux crustaux intermédiaires. Les études pétrologiques obtenues dans le mur et le toit de ce contact démontrent la juxtaposition, au moment du rifting, de ces roches de croûte supérieure coiffées de reliques de sédiments mésozoïques pré-rift à une portion de croûte inférieure localisée à des profondeurs de près de 30


1.
P-T-t path of the Malenco peridotites (pre-rift subcontinental mantle)
P-T-t path of the Malenco crust-mantle boundary (pre-rift Lower crust)
P-T-t path of the Bernina/Err intrusive (pre-rift upper crust)
P-T-t path of the Sondalo gabbro (pre-rift middle crust)
P-T-t path of the Lower Platta peridotites (pre-rift subcontinental mantle)
Triassic situation
Mesozoic cover
Upper crust (intruded by felsic material)
e.g. Bernina intrusive
Middle crust (intruded by felsic and gabbroic material)
e.g. Sondalo gabbro
Lower crust intruded by Permian Gabbros
e.g. Braccia gabbro
~30 km (crust-mantle boundary)
Lithospheric mantle
10 km
Err unit
Bernina unit
Ortler unit
Campo unit
Grosina unit
Malenco unit
Platta unit
Margna unit
Sella unit
Tonale unit
Languard unit
Grosina detachment
Eita shear zone
Engadine line
Insubric line
D1 thrust fault (eo-Alpine phase)
D2 Normal fault
Turba shear zone
Insubric,Engadine line and associated structures
movement zone
Jurassic low-angle detachment fault (late rifting)
Jurassic thinning structure (rifting)
2.
upper crust
middle crust
lower crust
Margna shear zone
e.g Pogallo type shear zone
Postrift sediments (Upper Jurassic-Lower cretaceous)
Middle Jurassic sediments (late rifting) "Calcareous" synrift
Lower Jurassic sediments (early rifting)
Triassic sediments
Felsic upper crust
Felsic middle/Lower crust
Acid/mafic intrusives
Subcontinentak mantle
Infiltrated subcontinental mantle
stretching faults
detachment fault/shear zone (brittle regime) faults
decollement (ductile layer)
exhumation faults
successively younger faulting systems

km (0,8 GPa), pendant le Permo-Trias (Müntener et al. 2000) (Fig. 2c). Par ailleurs des études précédentes ont proposé que la zone de cisaillement de Margna marquant la limite entre ces deux domaines crustaux était active pendant le rifting Jurassique (Hermann et Müntener 1996, Bissig et Hermann 1999). Ces travaux ont donc révélé, dans la marge distale Adriatique, l'amincissement de la croûte pendant le rifting de 20 à 25 km , avec l'absence de croûte moyenne (Fig 2).

1. 4. Les reliques de la marge Adriatique dans les Alpes du Sud

Des observations complémentaires peuvent être apportées par les reliques de la marge Adriatique dans les Alpes du Sud. Dans cette région, les marques de l'évolution de la marge proximale ont été remarquablement conservées dans le bassin Lombard montrant une évolution similaire à celle des nappes Austroalpines supérieures. (e.g. Bernoulli 1964, Bertotti 1993). En revanche, les reliques de la « necking zone » n'ont pas été identifiées, les affleurements n'étant pas favorables. Il est probable que cette zone intermédiaire soit située dans les unités de Biella et Gozzano High, unités qui se placent entre les domaines de marges proximale et distale. De nombreuses études ont démontré la juxtaposition plus à l'Ouest de croûte supérieure formant la zone Strona-Ceneri avec la croûte inférieure de la zone d'Ivrée-Verbano (e.g Handy 1987 ; Handy et al. 1999). Cette portion de croûte inférieure résidant à des pressions de 0,6-0,7 GPa a été exhumée pendant le rifting Jurassique (180-210 Ma). Dans la marge distale, l'exhumation à la fin du rifting des niveaux les plus profonds de la croûte est corroborée par l'identification de clastes de croûte inférieure dans les sédiments syn-rift de la zone du Canavese voisine de la zone d'Ivrée. La marge Adriatique préservée dans les Alpes du Sud montre donc une évolution et une architecture crustale semblable à celle des nappes Austroalpines. Cependant le manque d'accessibilité aux possibles reliques de la « necking zone » ne permet pas de répondre à toutes les questions posées dans cette région.

Fig. V. 2: (1) Carte géologique des nappes Austroalpines et Sud-Pénnique au Sud-Est de la Suisse et au Nord de l'Italie montrant la position des différentes unités pendant le Permien associée à une section crustale simplifiée montrant la position des différentes unités avant le rifting au Trias. Trajet P-T-t pour (a) La croûte moyenne pré-rift (gabbro de Sondalo, Tribuzio et al. 1999 ; Braga et al. 2001, 2003) ; (b) La croûte supérieure pré-rift (roches intrusives de Err et Bernina, Peters 2005, 2007) ; (c) La croûte inférieure pré-rift (la transition croûte-manteau de Malenco, Müntener et al. 2000 ; Villa et al. 2000) ; (d) Le manteau subcontinental supérieur pré-rift (unités de Malenco et Platta supérieure, Müntener et al. 2010 et références citées) ; (e) Le manteau subcontinental inférieur pré-rift (unité de Platta inférieure, Müntener et al. 2010 et références citées). (2) Reconstruction de la marge Adriatique à la fin du Jurassique montrant la position des différentes unités représentées sur la carte.

1. 5. Les reliques de la marge Adriatique dans les unités de Sesia-Dent blanche

Localement, les parties internes de l'orogène possèdent des unités d'affinités Adriatiques fortement déformées et métamorphisées par la compression Alpine (e.g. Schmid et al. 2004). Cependant, l'architecture complexe des zones distales de marge peut toujours être identifiée, notamment dans la zone de Sesia. Cette zone a été très fortement réactivée lors de l'orogène Alpine qui a rééquilibré les lithologies dans les faciès de haute pression. Une étude récente par Babist et al. (2006) a montré que la zone de Sesia est composée de l'imbrication d'allochtones extensionnels, composés majoritairement de reliques de croûte supérieure et/ou inférieure. Ces données viennent confirmer les observations précédentes montrant que les parties distales des marges sont composées de croûte supérieure et inférieure.

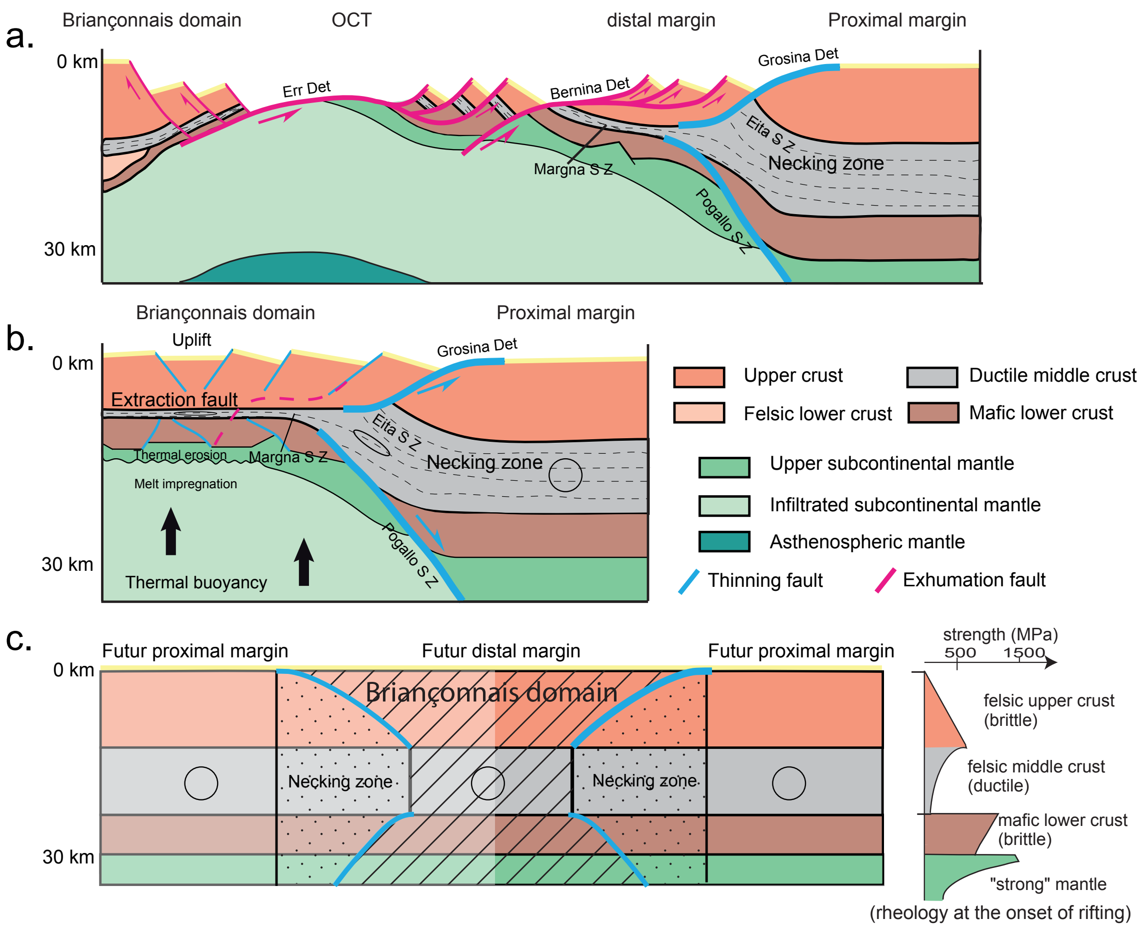


Fig. V. 3: Figure schématique montrant la déformation des différentes portions de la lithosphère en réponse au rifting. (a) Phase finale d'exhumation, (b) Phase d'amincissement, (c) Phase initiale pré-rift.

L'exhumation pendant le Jurassique des niveaux crustaux les plus profonds est mise en évidence par de nombreuses études récentes (Hermann et Müntener 1996 Handy et al. 1999, et références citées, Villa et al. 2000 ; Müntener et al. 2001). Elles démontrent la préservation d'une forte architecture crustale issue du rifting jurassique dans les parties distales des marges même si celles-ci ont été fortement réactivées lors de la collision. Antérieurement à ces travaux, de nombreux âges jurassiques obtenus par la méthode $^{40}Ar/^{39}Ar$ ont été mésestimés ou interprétés comme représentant un mélange entre l'orogenèse Varisque et Alpine. Il est intéressant de noter que la plupart de ces âges jurassiques obtenus par la méthode $^{40}Ar/^{39}Ar$ proviennent des parties distales des marges. Ceci peut être expliqué par le fait qu'à l'opposé des parties proximales qui montrent les marques d'une exhumation très limitée, les parties distales sont marquées par une forte exhumation des niveaux crustaux les plus profonds pendant le rifting. L'intensité de cette exhumation permet la fermeture des différents chronomètres utilisés par la méthode $^{40}Ar/^{39}Ar$. Ces âges peuvent être préservés au sein d'unité ayant subi un métamorphisme Alpin de haute pression tant que les roches n'ont pas été rééquilibrées.

Il a été établi, grâce aux données géophysiques sur les marges actuelles, que la croûte continentale est fortement amincie depuis le domaine de marge proximale et de « necking zone » vers le domaine distal. Les travaux menés à terre dans ce travail sur les marges fossiles permettent quant à eux de contraindre l'origine des roches ayant subi l'extension lithosphérique due au rifting. Cette extension se traduit par la juxtaposition latérale de domaines crustaux distincts qui étaient initialement superposés en une section verticale.

2. Les processus contrôlant l'extrême amincissement crustal entre les marges proximales à l'épaisseur crustale quasi normale, et les marges distales avec une épaisseur crustale voisine de 10 km *(Publications 2 et 3)*

Mon étude a permis la caractérisation de l'architecture présente dans les reliques d'une « necking zone » (unité de Campo/Grosina) et d'une marge distale (unité de Bernina) exposées dans la marge fossile Adriatique. Les travaux structuraux, pétrologiques et thermochronologiques ont permis d'établir l'évolution de ces domaines paléogéographiques depuis le Permien jusqu'à leurs implications dans l'orogène Alpine.

Des structures extensives accommodant l'extension lithosphérique sont décrites dans les reliques de la « necking zone » des unités de Campo/Grosina et de la marge distale préservée dans l'unité de Bernina. La caractérisation de ces structures est complétée par les données des travaux antérieurs sur les Alpes du Sud et plus particulièrement sur la zone d'Ivrée-Verbano. Ces structures extensives permettent l'amincissement lithosphérique dans le dommaine de la « necking zone » et accommodent l'amincissement crustal important de 30 à 10 km enregistré dans les parties distales des marges. Ces structures sont :

(1) des structures actives dans le domaine cassant considérées comme des détachements concaves (e.g. détachement de Grosina),

(2) des zones de cisaillement (e.g. zone de cisaillement d'Eita) à faible pendage localisées dans les niveaux crustaux intermédiaires,

(3) des zones de cisaillement à pendage modéré à faible permettant l'exhumation de croûte inférieure et sa juxtaposition avec les parties supérieures de croûte (e.g zone de cisaillement de Pogallo).

En complément, d'autres structures extensives sont décrites dans la partie distale de la marge. La zone de cisaillement de Margna en position distale n'est pas corrélable avec certitude aux structures de la « necking zone » précédemment décrites, cependant elle a joué un rôle important sur l'amincissement crustal.

2. 1. Structures extensives de la « necking zone »

Détachement de Grosina (Figs. 2 , 3)

La première structure importante est le détachement de Grosina. Ce détachement, localisé au toit de l'unité de croûte supérieure de Grosina, peut être suivi sur plus de 8 km ainsi qu'à l'interface entre la croûte moyenne exhumée de Campo et l'unité de Bernina comportant la croûte supérieure et la couverture sédimentaire mésozoïque. La déformation le long de cette structure s'effectue dans le domaine cassant indiqué par l'absence de recristallisation dynamique du quartz. Cette structure de détachement s'enracine dans la croûte moyenne (unité de Campo) ; elle permet ainsi l'exhumation de la crôute moyenne au sein de la « necking zone » et sa juxtaposition à la croûte supérieure de la marge distale. Cette faille de détachement à géométrie concave traverse les niveaux supérieurs et intermédiaires de la croûte, permet leur exhumation et marque la limite entre les domaines proximal et distal de la marge. Le détachement de Grosina représente donc une structure d'exhumation spécifique de la « necking zone », mais surtout une structure de découplage de la marge distale ou zone extrêmement amincie et de la marge proximale peu ou pas amincie.

<u>*Zone de cisaillement d'Eita et structures associées (Figs. 2, 3)*</u>

La zone de cisaillement d'Eita marque la limite entre l'unité de Campo représentant une croûte moyenne pré-rift et l'unité de Grosina relique d'une croûte plus superficielle. Cette zone est reconnue le long d'une même interface structurale au pendage relativement faible et peut être suive sur plusieurs dizaines de kilomètres. Elle est caractérisée par une forte déformation mylonitique rétrograde dans le faciès schiste vert à des températures entre 300 à 400°C enregistrées par les textures du quartz et l'absence de biotite. Le cisaillement d'Eita est associé à une foliation pénétrative gardant la même orientation dans les premiers 200 m de l'unité de Grosina qui enregistre des conditions métamorphiques similaires. Le fonctionnement de la zone de déformation d'Eita est daté par la méthode $^{40}Ar/^{39}Ar$ sur muscovite établissant un âge de refroidissement de 203,37±1,26 Ma.

La zone de cisaillement d'Eita semble ainsi être une structure efficace pour accommoder la réduction de la croûte moyenne pendant le rifting au niveau de la « necking zone ». Des structures similaires ont été identifiées dans la zone de Campo-Grosina, cependant leur âge reste incertain du fait du manque de données isotopiques. Ces zones mylonitiques d'épaisseur métrique généralement rétrogrades sont actives à des températures entre 500 et 300°C. Cet intervalle de température est trop « froid» pour correspondre aux critères du Permo-Trias et trop « chaud » pour relater la déformation Alpine dans la zone. Il est donc vraisemblable que ces structures aient été actives au cours du rifting.

Au toit de l'unité de Grosina, il est possible d'observer une zone cisaillante ayant enregistré des conditions métamorphiques similaires à celles d'Eita. Cette zone est recoupée par le détachement de Grosina avec une discordance angulaire de l'ordre 20 à 30°. Il est toutefois difficile de déterminer si cette zone représente la continuité du détachement de Grosina ou une structure similaire à la zone de cisaillement d'Eita.

La zone de cisaillement d'Eita et les structures analogues reconnues accommodent pendant le rifting l'amincissement des niveaux crustaux intermédiaires le long d'horizons tectoniques localisés. Ces structures créent un niveau de découplage dans la croûte moyenne ductile entre la croûte supérieure et la croûte inférieure qui ont un comportement plus « rigide ».

Ainsi, dans le modèle proposé dans cette étude, en cohérence avec les géométries similaires décrites dans les « metamorphic core complex » (Lister et Davis 1989), la zone d'Eita apparait être une structure faisant partie d'un système extensif et qui gouverne l'étirement des niveaux intermédiaires de la croûte.

Zone de cisaillement de Pogallo ou structure analogue (Figs. 2, 3)

Dans les reliques de la marge Adriatique exposées dans les nappes austroalpines, les parties les plus profondes de la « necking zone » ne sont malheureusement pas apparentes. Cependant, ces reliques peuvent être étudiées dans les Alpes du Sud, domaine aux évolutions similaires, afin de contraindre les processus d'amincissement dans les parties les plus profondes de la croûte. Dans cette région, les études précédentes ont caractérisé le contact entre la croûte supérieure pré-rift de Strona-Ceneri et la croûte inférieure pré-rift d'Ivrée-Verbano (Handy et al. 1999 et références citées). Le contact entre ces deux niveaux crustaux est une zone de déformation pouvant oscillant entre un à trois kilomètres de largeur, appelée la zone de cisaillement de Pogallo (Handy 1987). Contrairement à la zone d'Eita, qui se cantonne au même niveau crustal, la zone de cisaillement de Pogallo recoupe les niveaux de la croûte inférieure. Ceci est exprimé par les conditions métamorphiques le long du contact qui varient du faciès amphibolitique moyen (650°C) au faciès schiste vert supérieur (450°C). Les datations (méthode K/Ar sur biotite et muscovite et $^{40}Ar/^{39}Ar$ sur muscovite) effectuées à la fois sur le contact et dans le toit de la zone représentée par la zone d'Ivrée-Verbano indiquent une exhumation réalisée entre 180 à 210 Ma (Handy et al. 1999). Ces résultats ont été complétés plus récemment par la datation, par la méthode U-Pb, de la cristallisation d'une titanite indiquant un âge de 173±4 Ma pour le métamorphisme rétrograde associé à l'amincissement crustal le long de la zone de Pogallo (Mulch et al. 2002). Ainsi, ces données révèlent que la structure cissaillante de Pogallo a été active durant le rifting Jurassique. Handy (1987) avait établi que cette structure permettait l'exhumation de croûte inférieure pré-rift jusqu'à des profondeurs voisines de 10 km, recoupant la croûte continentale et permettant sa juxtaposition avec la croûte supérieure durant le rifting.

A partir de la reconstruction de la géométrie initiale de cette structure de cisaillement, il a été montré que son pendage était compris entre 10 et 34° pendant le rifting (Handy et al. 1987). Ces valeurs sont similaires à celles du pendage du Moho sous la « necking zone » dans les marges actuelles.

Ainsi, la zone de cisaillement de Pogallo peut être interprétée comme la structure accommodant l'exhumation de croûte inférieure dans la « necking zone » et cela jusqu'à des profondeurs voisines de 10 km.

2. 2. Structure extensive dans la marge distale

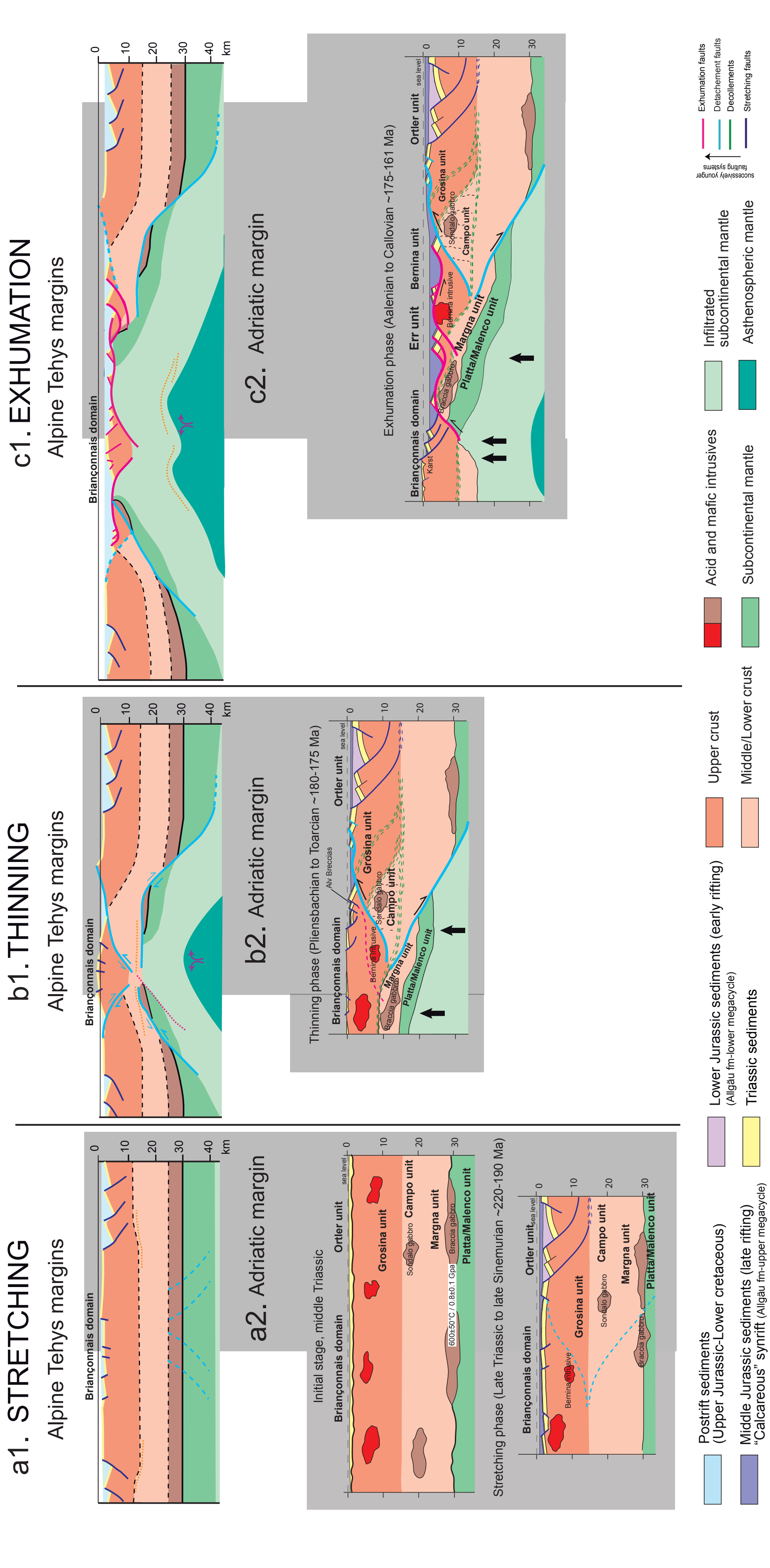
Zone de cisaillement de Margna, une faille d'extraction (Figs 2, 3)

Dans l'unité de Margna, dans les nappes Austroalpines représentant la marge distale Adriatique, une structure semblable aux précedentes a été reconnue et appelée la zone de cisaillement de Margna. Elle juxtapose une croûte supérieure préservant sa couverture mésozoique à une croûte inférieure qui se trouvait à une profondeur de 30 km au Permien et au Trias. L'amincissement crustal accommodé par la zone de Margna est de 25 km (Hermann et Müntener 1996 ; Bissig et Hermann 1999). La relation entre cette zone de cisaillement et les zones précédemment décrites reste encore mal connue. Cependant elle pourrait représenter la continuation du découplage entre la croûte supérieure et la croûte inférieure de la « necking zone » dans le domaine distal. Cette faille pourrait représenter une structure semblable aux failles d'extraction proposées par Froitzheim et al. (2006).

3. Evolutions tectonique, spatiale et temporelle du rifting à l'échelle des marges passives peu magmatiques *(Publication 2, 3)*

Les mécanismes de l'extension lithosphérique

L'ensemble de ces observations va nous mener à proposer un modèle global rendant compte des mécanismes d'amincissement lithosphérique. Les critères déterminants du modèle, établis par ces structures d'amincissement localisées dans la marge fossile Adriatique sont intégrés aux données antérieures, notamment stratigraphiques et pétrographiques, réalisées sur le domaine Téthysien Alpin. Cependant pour permettre la compréhension globalisée et cohérente de l'amincissement crustal, il devient nécessaire de considérer la totalité du système et plus précisément de prendre en compte les données de la marge conjuguée (Fig. 4).


a1. STRETCHING
Alpine Tehys margins
Briançonnais domain
a2. Adriatic margin
Initial stage, middle Triassic
Briançonnais domain
Ortler unit
Grosina unit
Campo unit
Sondalo gabbro
Margna unit
Braccia gabbro
Platta/Malenco unit
600±50°C / 0.8±0.1 Gpa
Stretching phase (Late Triassic to late Sinemurian ~220-190 Ma)
Bernina intrusive
b1. THINNING
Alpine Tehys margins
b2. Adriatic margin
Thinning phase (Pliensbachian to Toarcian ~180-175 Ma)
Alv Breccias
c1. EXHUMATION
Alpine Tehys margins
c2. Adriatic margin
Exhumation phase (Aalenian to Callovian ~175-161 Ma)
Err unit
Bernina unit
Karst
sea level
km
Postrift sediments
(Upper Jurassic-Lower cretaceous)
Middle Jurassic sediments (late rifting)
"Calcareous" synrift (Allgäu fm-upper megacycle)
Lower Jurassic sediments (early rifting)
(Allgäu fm-lower megacycle)
Triassic sediments
Upper crust
Middle/Lower crust
Acid and mafic intrusives
Subcontinental mantle
Infiltrated subcontinental mantle
Asthenospheric mantle
successively younger faulting systems
Exhumation faults
Detachement faults
Decollements
Stretching faults

Le début du rifting : une déformation délocalisée sur l'ensemble du système Téthysien Alpin. (Fig. 4a)

De la fin du Trias (~200-220 Ma) jusqu'au à la fin du Sinémurien et le début du Pliensbachien (~185-190 Ma), la déformation extensive est concentrée sur l'ensemble du système des futures marges téthysiennes Alpines. Cette déformation se traduit par la création de bassins de rift contrôlés par des failles normales à fort pendage. Elle est localisée sur les deux futures marges et documentée par de très nombreux exemples dans le domaine Alpin (e.g. Bourg d'Oisans, Il Motto, Monte Generaso). L'épaisseur crustale sous ces bassins de rift n'est pas accessible dans le système Alpin. Cependant à partir des données provenant des marges actuelles, principalement les profils sismiques, l'hypothèse peut être faite que cette déformation n'a eu que très peu d'incidence sur l'épaisseur crustale (e.g bassin de Jeanne d'Arc).

L'amincissement crustal, un évènement fondamental pour l'acquisition de la géométrie des marges (Fig. 4b)

A partir de la fin Sinémurien et du début du Pleinsbachien, le domaine Alpin va subir une forte structuration générée par la création du rift. Cette période coïncide avec la migration de la déformation des parties proximales aux parties distales de la marge. Cette migration est caractérisée par la création de bassins de rift, contrôlés par des failles normales et une subsidence importante, dans des domaines qui étaient jusqu'à lors épargnés par une tectonique extensive.

Peu après le début de cette migration, le domaine Briançonnais conjugué à la marge Adriatique va subir une évolution atypique. En effet à partir du Toarcien-Aalenien (~170-185 Ma) le domaine Briançonnais est affecté par un soulèvement important qui va le porter jusqu'aux conditions subaériennes. L'évolution atypique du bloc Briançonnais est déconnectée de l'évolution du domaine Téthysien qui est marqué à l'opposé par une subsidence forte. L'individualisation de ce bloc continental est accompagnée par l'exhumation de différents niveaux de la croûte continentale. En effet, les nombreuses datations et études structurales effectuées dans les parties distales des marges du domaine Téthysien tendent à montrer que les niveaux profonds de la croûte ont subi une exhumation à partir de 180-185 Ma. Il convient de préciser qu'à cette période les « necking zones » Européenne et Adriatique marquent les limites du bloc Briançonnais. Les exhumations sont accommodées par les structures décrites précédemment (e.g. détachement de Grosina, zone de cisaillement d'Eita, zone de cisaillement de Pogallo).

Fig. V. 4: Modèle conceptuel montrant l'évolution du domaine Téthysien Alpin pendant l'amincissement lithosphérique et plus particulièrement la marge Adriatique. (a1) La phase initiale du rifting, (a2) Stade antérieur au rift et phase initiale d'amincissement de la marge Adriatique. (b1) Stade d'amincissement lithosphérique, (b2) phase d'amincissement dans la marge Adriatique. (c1) Stade final du rifting avec exhumation du manteau subcontinental et création des bassins Valsain et Liguro-Piémontais, (c2) phase d'exhumation dans la marge Adriatique et dans le bassin Liguro-Piémontais. Il est important de noter l'évolution du bloc Briançonnais pendant le rifting.

Le détachement de Grosina et la zone de cisaillement de Pogallo vont conduire à l'amincissement lithosphérique ainsi qu'à l'exhumation de niveaux crustaux plus profonds. Ces évolutions sont accompagnées par l'établissement de niveaux de découplages dans la croûte moyenne (e.g zone de cisaillement d'Eita). Si ces structures ont été décrites principalement dans la marge Adriatique, il est fort probable néanmoins qu'elles furent aussi actives dans la marge conjuguée Européenne. Elles vont engendrer l'individualisation et le découplage tectonique du domaine Briançonnais des autres domaines des marges proximales Européenne et Adriatique.

L'individualisation de ce bloc va s'accompagner d'un soulèvement tectonique mettant ce domaine dans des conditions subaériennes durant le Toarcien-Aalénien jusqu'au Bathonien-Callovien. Cette évolution isostatique atypique démontre que l'amincissement crustal précédemment décrit est couplé avec un amincissement lithosphérique. Ce dernier est documenté dans le domaine de la Téthys Alpine par l'imprégnation par du magma du manteau subcontinental lors probablement de la phase d'amincissement (~180 Ma) (e.g Piccardo 2008 ; Manatschal et Müntener 2009 ; Müntener et al. 2010 et références citées, manteau de Platta inférieure (Fig. 2e)). Cette imprégnation du manteau apporte un support thermique au bloc Briançonnais lui permettant de rester en milieu subaérien durant le rifting (Fig. 3).

<u>L'exhumation de manteau subcontinental et la création des marges distales : le stade final du rifting (Fig. 4c)</u>

La phase finale du rifting va se faire aux dépens du bloc Briançonnais. En effet ce bloc découplé lors de l'amincissement crustal va être délaminé par des failles de détachement à faible pendage (e.g. détachement de Bernina). La délamination va créer des marges distales de part et d'autre. Les domaines nouvellement créés sont caractérisés par une épaisseur crustale réduite (<10 km, cf. marges actuelles). Cette interprétation est cohérente avec les descriptions stratigraphiques effectuées dans le domaine Briançonnais établissant la similitude des faciès pré-rift avec ceux des parties distales de la marge Adriatique. De plus, dans cette dernière, la couverture mésozoïque et localement une partie de la croûte supérieure reposent sur le socle sous forme d'allochtones extensionnels reliés à une faille de détachement qui est caractérisée par un sens de cisaillement corroborant la délamination du Briançonnais.

La dernière phase d'extension accommodée par des failles de détachement à faible pendage va permettre d'amplifier l'amincissement crustal de 10 à 0 km et finalement provoquer l'exhumation de manteau subcontinental. La délamination de part et d'autre du bloc Briançonnais (e.g marge Européenne, Adriatique) forme deux bassins sédimentaires reconnus dans les Alpes comme les domaines Valaisan et Liguro-Piémontais. Le domaine Liguro-Piémontais va évoluer vers l'océanisation et cette évolution distincte provoque une asymétrie des marges reliée à l'extension du bloc Briançonnais. La fin du rifting dans les Alpes est datée par l'intrusion de gabbro à 165 et 157 Ma (méthode U/Pb sur Zircons) (Bill et al. 1997 ; Schaltegger et al. 2002) ainsi que par le dépôt de sédiments siliceux datés du Bathonien-Oxfordien reposant sur le manteau exhumé et probablement sur la croûte océanique, qui représentent ainsi les premiers sédiments post-rift (Baumgartner 1987). C'est à cette même période que le bloc Briançonnais va subir une rapide subsidence.

En conclusion, le rifting dans les marges fossiles peu magmatiques préservées dans les Alpes est caractérisé par une succession de structures extensives jouant le rôle de découpleurs qui vont permettre l'individualisation du bloc Briançonnais et le découplage entre la croûte supérieure et inférieure accommodé par la croûte moyenne réalisé au niveau la « necking zone ». Dans les Alpes, l'amincissement crustal enregistré à partir de la fin du Sinémurien et le début du Pliensbachien (180-185 Ma) aura des conséquences majeures sur l'architecture crustale et sur la structuration du domaine Téthysien en général.

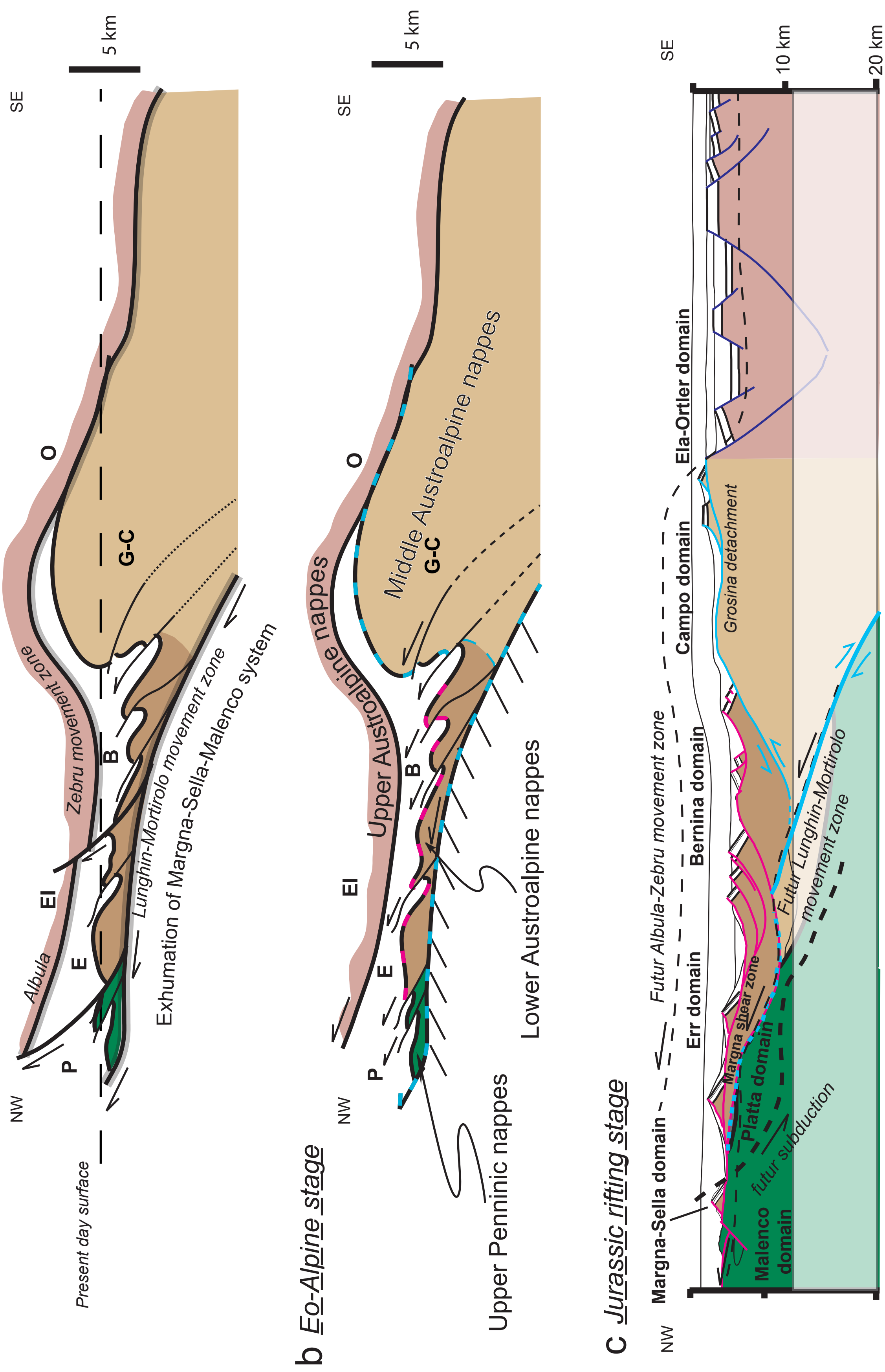
4. Réactivations des structures de rifting pendant la phase compressive alpine et impact de ces structures issues du rifting sur la géométrie des structures alpines *(Publication 1)*

Il a été établi que les nappes Austroalpines du Sud-Est de la Suisse et du Nord de l'Italie préservent les reliques des structures issues du rifting. Cette étude a montré que cette pré-structuration, cette géométrie complexe issue du rifting a des conséquences majeures sur les processus de réactivation pendant l'orogénèse Alpine. Je propose de prendre l'exemple de cette zone pour expliciter le rôle de la pré-structuration du rift sur l'architecture des nappes Austroalpines. Cette pré-structuration peut être considérée à différentes échelles.

L'architecture de la marge contrôlant la géométrie et la déformation Alpine (Fig. 5)

L'agencement des nappes à grande échelle va être contrôlé par l'architecture globale de la marge, telles que les variations d'épaisseur crustale, l'orientation des structures du rift par rapport à la direction de compression, etc…. Ce contrôle est visible dans le cas étudié des nappes Austroalpines et Sud Pénninique. Il est démontré que certaines zones de faiblesse, issues du rifting, vont être réutilisées tout long de l'histoire compressive Alpine. J'ai pu définir et caractériser deux zones de faiblesses majeures réactivées plusieurs fois durant les phases de déformations Austroalpines.

La première zone est celle d'Albula-Zebru qui représente le chevauchement des nappes Austroalpines supérieures sur les nappes Austroalpines inférieures. Cette structure majeure correspond au chevauchement de l'empilement sédimentaire de la marge proximale sur la marge distale. Le promontoire représenté par la « necking zone » a fortement influencé la géométrie


a D2 Meso-Alpine stage
NW
SE
5 km
Present day surface
Albula
P
E
El
B
G-C
O
Zebru movement zone
Lunghin-Mortirolo movement zone
Exhumation of Margna-Sella-Malenco system
b Eo-Alpine stage
NW
SE
5 km
Upper Austroalpine nappes
Middle Austroalpine nappes
Lower Austroalpine nappes
Upper Penninic nappes
C Jurassic rifting stage
NW
SE
10 km
20 km
Margna-Sella domain
Malenco domain
Platta domain
Margna shear zone
futur subduction
Err domain
Futur Albula-Zebru movement zone
Bernina domain
Futur Lunghin-Mortirolo movement zone
Campo domain
Grosina detachment
Ela-Ortler domain

de cette zone de chevauchement expliquant ainsi la géométrie de l'empilement des nappes et l'absence de couverture sédimentaire préservée dans les domaines conservant les reliques de la « necking zone ».

La deuxième zone décrite est celle de Lunghin-Mortirolo. Cette structure est essentielle pour comprendre les mécanismes de réactivation et même éventuellement l'initiation de la subduction. Elle marque la limite entre deux aires paléogéographiques semblables représentant une marge distale mais ayant subi une évolution Alpine distincte. Les unités d'Err-Platta répresentant respectivement la marge distale et la TOC (la zone de manteau exhumé) n'ont été que très peu affectées par la compression au Crétacé et au Tertaire. Les chevauchements sont limités à quelques zones de déformation localisées et le métamorphisme n'a jamais été supérieur au faciès schiste vert à amphibolite à épidote. En revanche, les unités de Margna/Sella-Malenco associées au même domaine paléogéographique ont été très fortement affectées par la déformation Alpine associée à une déformation pénétrative et une recristallisation métamorphique de faciès schiste vert. Une hypothèse expliquant cette évolution distincte est de relier celle-ci au début du processus de subduction de la Téthys Alpine. Cette interprétation a déjà été proposée par Froitzheim et al. (1996). La zone de Margna-Sella, relique d'une croûte continentale avec sa couverture mésozoïque, est entourée de deux zones distinctes de manteau subcontinental exhumé lors du rifting, (les unités de Malenco et de Platta). Cette géométrie a permis à certains auteurs de proposer que l'unité Margna-Sella est un allochtone extensionnel composé de croûte continentale et de sédiments mésozoïques, reposant sur le manteau exhumé. Ces résultats signifient que la subduction (fermeture de l'océan Liguro-Piémontais) peut localement s'initier dans des fenêtres de manteau subcontinental exhumé, en entraînant la subduction de blocs continentaux allochtones (Fig. 5c).

Les nouvelles observations issues de ce travail permettent de suivre, dans toute la zone considérée, la limite entre les domaines peu ou pas affectés et les domaines fortement réactivés par la phase alpine. De cette étude cartographique, il découle que la zone dite de Lunghin-Mortirolo représentait certainement déjà une limite lithologique majeure au moment du rifting Jurassique, qui a été ensuite réactivée à plusieurs reprises par la déformation Alpine.

Fig. V. 5: Reconstruction tectonique des nappes Austroalpines et Sud-Pénnique dans le Sud-Est de la Suisse et le Nord de l'Italie. (a) D2 phase méso-Alpine avec exhumation des unités de Margna-Sella-Malenco. (b) Réactivation des structures issues du rifting pendant la phase D1 éo-Alpine. Il est important de noter le contrôle de l'architecture du rift sur la géométrie des chevauchements. (c) Architecture de la marge Adriatique à la fin du Jurassique. Initiation de la subduction dans la fenêtre de manteau subcontinental est représentée. Cette géométrie explique la déformation Alpine plus forte subie par les unités de Margna-Sella représentant un allochtone extensionnel sur le détachement Jurassique. Il est important de noter le rôle majeur des zones de Lunghin-Mortirolo et Albula-Zebru qui représentent des discontinuités issues du rifting et qui vont être réutilisées tout au long de la déformation Alpine.

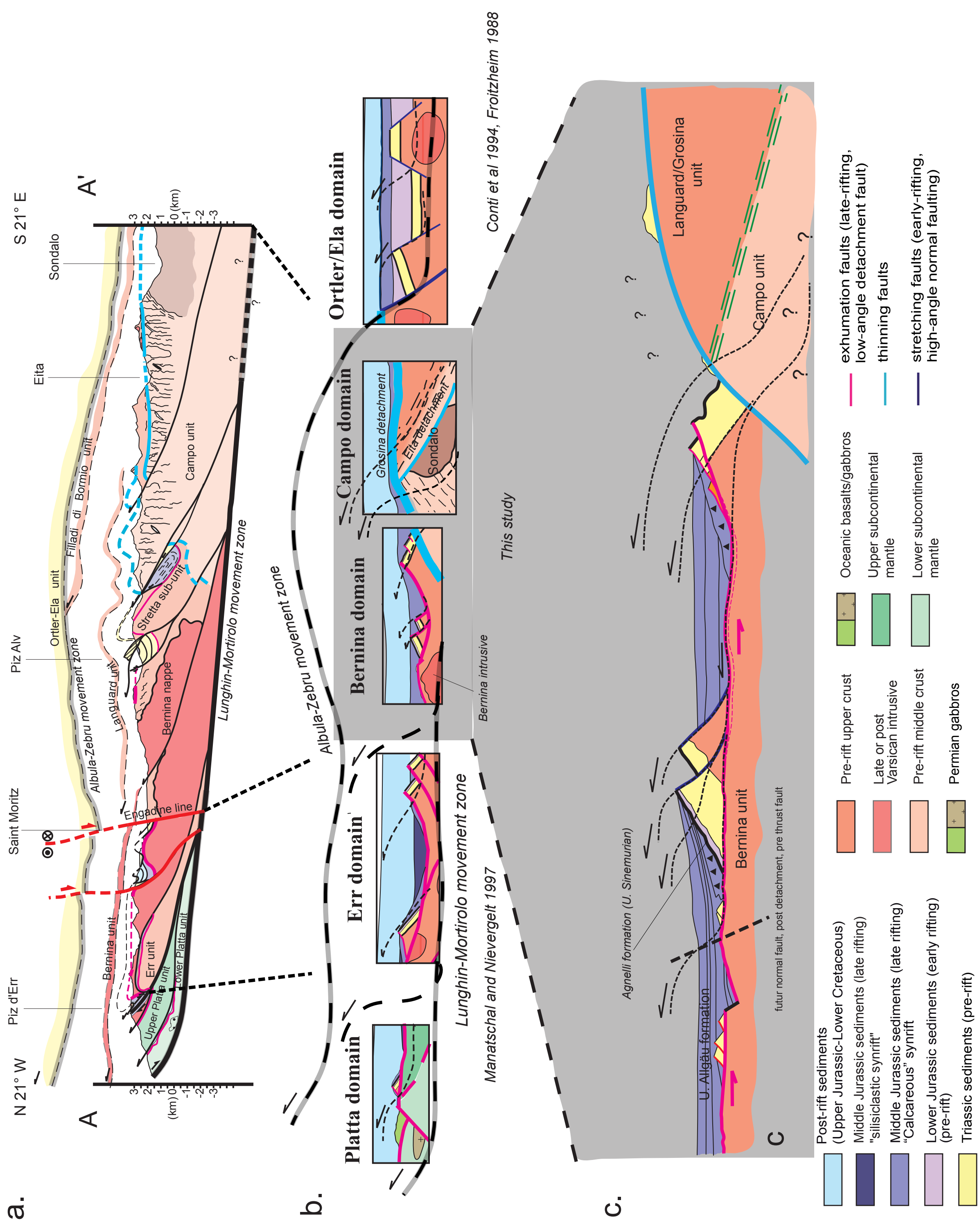


Fig. V. 6: (a) Coupe Alpine (A-A') à travers les nappes Austroalpines et Sud-Pénnique dans le Sud-Est de la Suisse et le Nord de l'Italie, la localisation de la coupe sur trouve sur la figure 2. (b) Reconstruction de la géométrie des nappes Austroalpines et Sud-Pénnique. Il est important de noter le rôle de l'architecture et de la marge distale et de la necking zone sur la réactivation Alpine. (c) Reconstruction de l'unité de Bernina représentant une marge distale et les unités de Campo-Grosina, reliques d'une necking zone. Le tracé des futurs chevauchements est représenté.

Ainsi l'architecture acquise de la marge contrôlera la géométrie des chevauchements ultérieurs et cet héritage très marqué de la structuration des marges sont des zones de faiblesses majeures qui seront réutilisées efficacement lors de la réactivation Alpine.

Les chevauchements Alpins, des structures préexistantes inversées et réactives (Fig. 6)

Les structures d'échelles kilométriques, parfois moindres, telles que les blocs basculés ou allochtones vont contrôler les processus de réactivation à de plus petites échelles. En effet en fonction de leurs orientations face à la compression Alpine et en fonction de la direction des chevauchements, les failles normales à fort pendage vont contrôler les géométries et le style de déformation. Cette réactivation est fortement reliée au style tectonique de ces zones.

Les unités Austroalpines supérieures possédant les reliques de la marge proximale sont caractérisées par une couverture mésozoïque continue. La réactivation y est marquée principalement par l'inversion des bassins préexistants en blocs basculés créés lors du rifting Jurassique. Dans la zone d'étude, les directions d'extension Jurassique étant parallèles aux directions de compression Alpine, la réactivation se traduira principalement de deux manières.

(1) Lorsque l'orientation des failles normales Jurassiques est antithétique au mouvement, les failles peuvent localement servir de rampe frontale pour le chevauchement Alpin. Cette géométrie est variable en fonction du pendage des paléo-failles normales qui peuvent être localement trop pentées pour être réactivées par des chevauchements.

(2) Si les deux structures sont de directions semblables, le chevauchement Alpin peut éventuellement décoller le bloc faillé Jurassique. Le niveau de décollement de ces chevauchements est alors localisé selon des horizons stratigraphiques précis, soit le Trias moyen dans le niveau de gypse, soit à la limite entre la couverture Permo-Triassique et le socle.

Il est à noter que la réactivation de la marge proximale Adriatique est localement distincte de celle de son homologue Européenne. Cette différence peut résulter de nombreux facteurs. Il est important de noter que la marge proximale Européenne est caractérisée par une faible épaisseur de sédiments pré-rift (~100 m) tandis que la marge Adriatique possède une couverture sédimentaire de plus d'un kilomètre. Cette différence peut expliquer en partie que dans les parties européennes proximales, le socle sot beaucoup plus impliqué dans la déformation alpine.

En revanche, la réactivation alpine est différemment enregistrée dans les parties distales et la « necking zone », étudiées dans cette thèse. Dans les « necking zones » et les parties distales des marges, les structures extensives impliquées sont marquées par la présence de failles normales à fort et faible pendage. Elles ont pour effet l'exhumation de roches crustales et la dislocation de la couverture pré-rift sous forme d'allochtones extensionnels localisés au toit du socle. Cette géométrie héritée du rifting va déterminer les zones de faiblesses réactivées pendant la compression Alpine. Les chevauchements Alpins vont ainsi réactiver principalement les détachements à faible pendage devenant des niveaux de décollement (Fig. 6). Comme les parties proximales des marges, les blocs basculés préservés en allochtones extensionnels vont soit servir de rampes frontales, soit être décollés et transportés par les chevauchements alpins. Ces deux styles différents de déformation sont observés dans la zone d'étude (Fig. 6).

Les reconstructions Alpines précédemment proposées ont pour hypothèse la continuité de la séquence pré-rift. Si ce postulat est applicable aux parties proximales des marges, il ne semble en revanche plus valide pour les parties distales. En effet, l'unité de Bernina étudiée est constituée d'écailles de sédiments pré-rift reposant sur une faille de détachement Jurassique. Cette dernière a provoqué la dislocation de la plateforme pré-rift ainsi que l'exhumation de croûte supérieure. Ainsi la répartition actuelle entre croûte continentale, sédiments pré-, syn-, et post-rift est dépendante de la géométrie Jurassique. Ceci est en accord avec la mise en évidence des contacts primaires des sédiments syn-rift reposant sur le socle paléozoïque avec un onlap de 15 à 20° et qui révélent l'exhumation de socle par le jeu des failles de détachement, puis le scellement par les sédiments syn-rift. L'existence de la discontinuité de la plate forme Jurassique dans les parties distales des marges est un élément clé de la compréhension des évolutions et des reconstructions Alpines.

5. Le rôle de l'héritage structural sur le contrôle du rift ainsi que sur l'orogène Alpine : Implications d'une meilleure connaissance des processus d'extension lithosphérique *(Publication 1, 3)*

La complexité structurale héritée du rifting conditionne la géométrie de l'orogène Alpine. Les variations de l'épaisseur crustale acquises lors de l'extension lithosphérique entre les zones proximales peu amincies, la « necking zone » et les parties distales de la marge où l'épaisseur est voisine de 10 km, vont induire une forte structuration de la marge dans son ensemble (Fig. 7). La variation de l'épaisseur de la croûte peut être reliée aux différences de l'intensité de la déformation, des conditions métamorphiques, du style tectonique lors de l'implication dans l'orogène Alpine des différents domaines de marge (Fig. 7). Les parties externes de l'orogène représentent principalement les reliques des marges proximales. Ces domaines sont marqués par une déformation relativement faible associée à des conditions métamorphiques ne dépassant pas le faciès schiste vert. Par similitude avec les caractéristiques des marges actuelles, il est raisonnable de penser que lors de leur implication dans la collision Alpine ces domaines

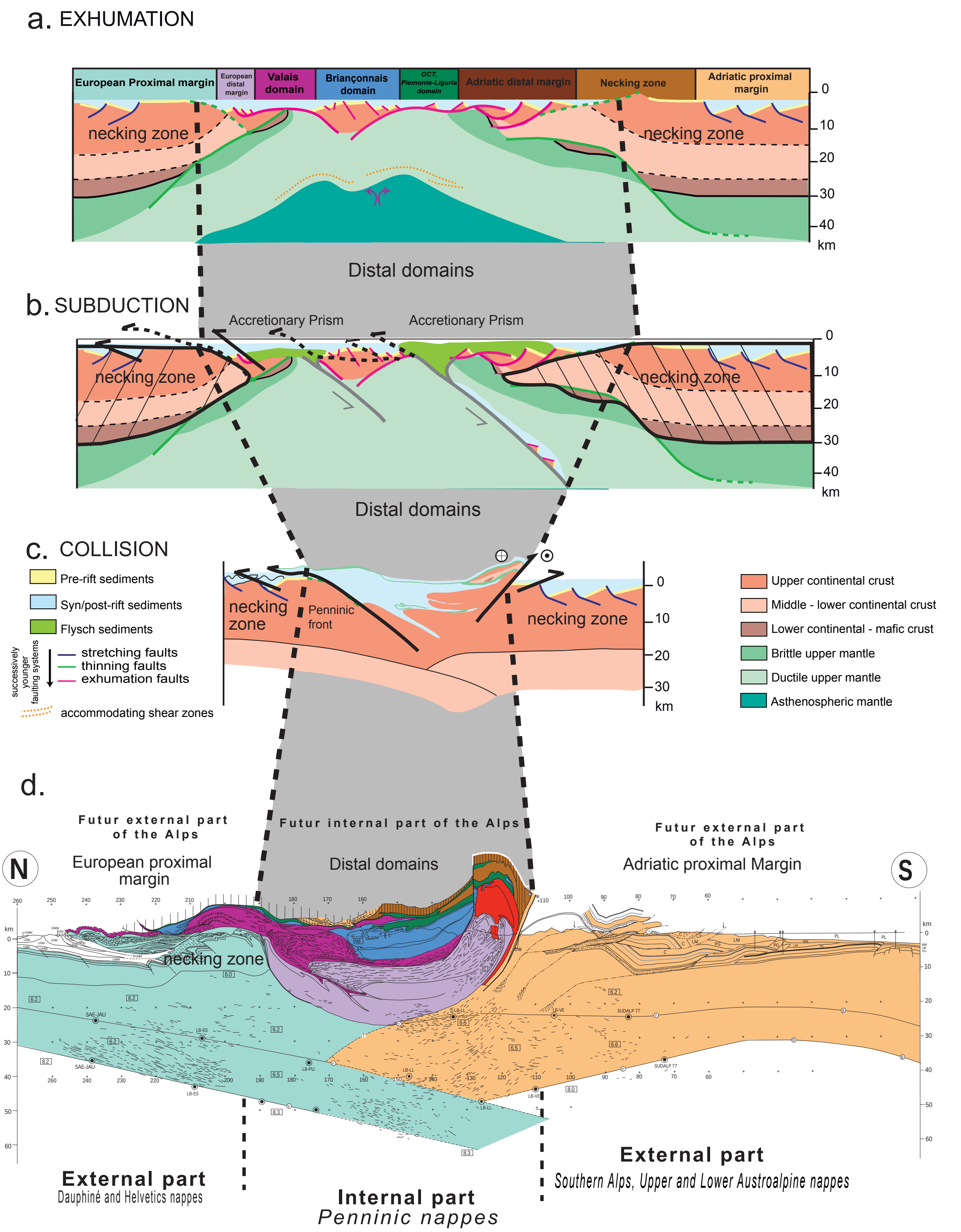


Fig. V. 7: Figure schématique montrant, à l'échelle du domaine Téthysien Alpin le rôle essentiel de l'architecture des marges sur la réactivation Alpine. (a) Situation des marges Téthysienne Alpine à la fin du Jurassique. (b) Situation lors du début de la subduction. (c) Situation lors de la collision continentale. (d) Coupe sismique NFP 20 à travers la chaîne Alpine (Schmid et al. 1996) montrant l'architecture actuelle des Alpes. Il est important de noter le rôle des deux necking zone, qui vont jouer le rôle de butoir. Celles-ci marquent la limite entre les parties externes représentant les marges proximales et les parties internes composées des reliques des parties distales.

possédaient toujours une épaisseur crustale voisine de 30 km. Ils n'ont été que très peu affectés par la collision Alpine, et au premier ordre cette déformation est limitée à la compression des bassins du rift Jurassique créés par failles normales (e.g. le bassin de Bourg d'Oisans). En revanche, les parties distales des marges trouvées actuellement dans les parties internes des Alpes étaient quant à elles caractérisées par une épaisseur crustale fortement réduite au cours du rifting. La déformation Alpine affectant ces domaines de croûte amincie est très différente de celle des parties externes (Fig. 7). En effet, les parties internes sont marquées une superposition métamorphique très forte atteignant les faciès de haute pression, avec une déformation Alpine polyphasée et pénétrative. Grâce aux données sismiques sur la chaîne Alpine, notamment sur la coupe NFP 20 (Schmid et al. 1996) dans les Alpes Centrales, il est possible de visualiser que l'essentiel des processus de subduction et de collision continentale est accommodé par les parties distales des marges. Ainsi, pendant les processus de fermeture et collision du bassin Téthysien, les « necking zones » vont jouer le rôle de butoir pour la subduction et vont placer les limites du prisme de collision (Fig. 7). La structuration Alpine entre domaine interne et externe est donc à relier pour partie à l'architecture crustale héritée du rifting.

Les zones externes des Alpes sont généralement bien contraintes structuralement et la part entre pré-structuration et déformation Alpine établie. Cependant, dans les zones internes de l'orogène, cette distinction est beaucoup plus difficile à définir. Il est donc essentiel pour établir des reconstructions paléogéographiques cohérentes de ces domaines d'acquérir les données relatives à l'architecture et la géométrie de la marge antérieurement à son implication dans l'orogène Alpine. Au cours des dernières années, de nombreuses études ont reconnu la persistance de structures ou d'âges Jurassiques reliés au rifting, et cela même dans des domaines ayant subi un métamorphisme de haut dégré, indiquant ainsi l'implication d'une pré-structuration forte (e.g. Babist et al. 2006). Il a été établi que des contacts majeurs Alpins représentaient des discontinuités majeures déjà présentes au cours du rifting. Cette pré-structuration a été montrée dans la zone de Sesia dans les Alpes Centrales par (Babist et al. 2006) ainsi qu'à la limite entre la zone de Zermatt-Sass et la zone du Combin dans les Alpes de l'Ouest. Dans le dernier cas, malgré un métamorphisme poussé jusqu'au faciès éclogitique pendant la fermeture du domaine océanique, il a été possible de reconnaître des structures pré-Alpines (Beltrando et al. 2010). Ces structures réactivées pendant la compression représentaient déjà une discontinuité majeure active lors du rifting.

Ces nouvelles études démontrent que pour reconstruire de façon cohérente (structures, âges…) la géométrie des parties Alpines les plus déformées, la prise en compte de l'architecture (structure, lithologie crustale, rhéologie…) des différents domaines de marge héritée au cours rifting est essentielle.

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CONCLUSION

Cette thèse a pour but de comprendre les mécanismes d'amincissement de la lithosphère lors du rifting dans les marges passives peu magmatiques. Les études précédentes s'étaient majoritairement focalisées : (1) sur les parties proximales des marges caractérisées par une croûte relativement peu ou pas amincie (~30 km) (2) sur les parties distales des marges montrant une croûte hyper amincie (~10 km) allant jusqu'à l'exhumation de manteau subcontinental dans les TOC. Les données provenant des marges passives actuelles ont permis de montrer que l'exhumation de manteau subcontinental et de croûte continentale dans les marges distales ne représente que la phase finale d'amincissement amenant l'épaisseur crustale de 10 à 0 km. Cependant l'extrême amincissement lithosphérique qui va amincir la croûte de 30 à 10 km reste méconnu. Cet amincissement dans les marges actuelles s'effectue dans la «necking zone» à la frontière entre marge proximale et distale. Les données provenant des marges actuelles ne permettent pas de répondre à ces questions dues aux manques de sismique réflexion et réfraction associés aux nombres insuffisants de forages. Dans le but de caractériser cet amincissement crustal notre étude s'est focalisée sur les marges fossiles de la Téthys Alpine préservées sur les Alpes de l'Ouest et Centrale. Plus particulièrement cette thèse s'est focalisée dans un premier temps dans les nappes Austroalpines du Sud-Est de la Suisse et du Nord de l'Italie avant d'intégrer dans un deuxième temps des observations et des données sur l'ensemble du système Alpin. Cette étude s'est basée sur un travail de cartographie et de géologie structurale couplé à des datations $^{40}Ar/^{39}Ar$ pour : (1) identifier les différents événements tectoniques issus de la compression Alpine, (2) reconstruire la géométrie d'une «necking zone» et d'une marge distale pendant le rifting, (3) identifier et décrire les structures extensives pouvant accommoder cet étirement lithosphérique, (4) intégrer ces données à l'échelle d'une marge passive peu magmatique ainsi que les implications pour la compréhension de l'orogène.

Les principaux résultats de ce travail sont :

- La reconnaissance et la description des différents événements tectoniques affectant les nappes Austroalpine et Sud-Pénnique dans le Sud-Est et de la Suisse et le Nord de l'Italie. Cette étude est plus particulièrement centrée sur les unités de Bernina-Campo-Grosina localisée à l'interface entre les reliques des marges distales (e.g. Err) et proximales (e.g. Ortler).
- La rétro-déformation et la reconstruction des unités de Bernina-Campo-Grosina, identifiant la zone de Bernina comme une relique de marge distale tandis que les unités de Campo-Grosina représentent une «necking zone». Cette étude a permis de mettre en avant le rôle primordial de l'héritage structural ainsi que l'influence de la géométrie issue du rifting lors de la réactivation Alpine.

- L'étude des reliques de la «necking zone» de Campo-Grosina indique que celles-ci se composent de lithologies qui se trouvaient dans la croûte supérieure à moyenne lors du Permo-Trias avant le rifting Jurassique. Des nouvelles données thermochronologiques ont permis de démontrer l'exhumation de cette croûte moyenne pré-rift lors du rifting (184-188 Ma) dans l'unité de Campo.
- L'identification des structures extensives accommodant l'amincissement lithosphérique Jurassique dans les reliques de la marge Adriatique. Il a été montré que la necking zone se caractérise par un système couplé de détachement d'échelle crustale (e.g. détachement de Grosina et zone de cisaillement de type Pogallo) accommodant respectivement l'exhumation de croûte moyenne à supérieure ainsi que l'exhumation de croûte inférieure.
- L'importance des niveaux de la croûtes intermédiaires dans la «necking zone». En effet ces niveaux vont jouer le rôle majeur de découplage entre croûte supérieure et inférieure. Ce découplage va être accommodé le long de zone de décollement (e.g. zone de cisaillement d'Eita)
- La caractérisation de l'architecture crustale de la marge distale Adriatique dans les unités de Bernina-Margna. Dans ces domaines, la croûte supérieure est directement juxtaposée à la croûte inférieure avec l'omission de la croûte moyenne. L'interface entre ces deux structures est caractérisée par une zone de cisaillement Jurassique (e.g. zone de cisaillement de Margna). Cette structure est interprétée comme une faille « d'extraction ».
- La caractérisation de l'architecture crustale de la necking zone ainsi que des parties distales de la marge Adriatique ont permis de proposer un modèle cohérent expliquant les processus ainsi que les structures permettant l'amincissement lithosphérique pendant le rifting.

Ces résultats ont été couplés à l'ensemble de la Téthys Alpine permettant ainsi la compréhension de l'évolution des marges passives dans leur ensemble. Cette thèse a permis d'établir un modèle cohérent pour expliquer l'évolution des marges passives peu magmatiques lors de l'extrême amincissement lithosphérique.

Cette étude a montré qu'une étroite interaction entre le système extensif lié au rifting ainsi que le système compressif lié à l'établissement de l'orogène est essentielle pour accéder à des reconstructions valides de ces deux systèmes. Les complexités issues du rifting ne peuvent être ignorées pour la compréhension de la réactivation et de l'imbrication des marges passives dans l'orogène.

Les nouvelles données et observations issues de ce travail conviendraient d'être couplées avec des systèmes actuels pour ainsi mieux contraindre l'évolution des marges passives peu magmatiques. En effet dans les marges actuelles, l'identification des structures ainsi que de l'architecture crustale reste problématique à cause du manque de données sismiques réflexion et réfraction et de forages.

En outre, la caractérisation de la complexité des marges passives de la Téthys Alpine peut permettre une interprétation plus cohérente au regard de la structure des marges de certaines unités d'affinités paléogéographiques mal définies.

Enfin, une perspective majeure de ce travail est la compréhension de la rhéologie de la croûte continentale lors de l'extension lithosphérique. Cette caractérisation permettra une meilleure calibration des modèles numériques relatifs à l'amincissement lithosphérique.

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Annexes

Annexe A – Méthode Ar-Ar : principe, méthode et applications.

Introduction :

Comme la méthode K-Ar, la méthode Ar-Ar est basée sur la désintégration du potassium 40 en argon 40, le rapport entre ces deux éléments étant proportionnel à l'âge de fermeture du système isotopique. Si la première méthode nécessite la mesure de deux aliquots, l'un pour le potassium et l'autre pour l'argon, la méthode Ar-Ar permet à l'aide d'une seule mesure d'obtenir l'«age» de l'échantillon, ce qui diminue d'autant les incertitudes de mesures.

Principe :

Le principe de la méthode est basé sur une idée originale : on irradie dans un réacteur nucléaire l'échantillon et un standard d'âge connu. On permet ainsi la formation d'un isotope artificiel, l'argon 39, à partir du potassium 40 naturel.

Le rendement de l'irradiation étant connu, on peut dès lors, à l'aide de la mesure d'un seul rapport isotopique et en faisant les corrections nécessaires, obtenir l'« âge » de l'échantillon.

Le second avantage de la méthode Ar-Ar est de donner non seulement un age global mais aussi un spectre d'âges grâce un à processus de dégazage progressif. Au lieu d'être fondu totalement comme c'est le cas pour la méthode K-Ar, on augmente la température graduellement par palier successifs («stepwise heating» [Turner, 1966]). En mesurant le rapport isotopique à chaque incrément, on obtient ainsi une succession d'âges jusque la fusion du matériel.

Signification et interprétation des spectres d'ages :

L'age « apparent » obtenu par la méthode K-Ar correspond à l'age de fermeture du système si et seulement si tout l'argon libéré lors du chauffage provient de la même génération de cristaux et que le système est resté fermé depuis sa formation. Pour cela, le système doit avoir totalement dégazé lors de l'évènement que l'on veut dater et n'avoir subi aucune crise occasionnant des pertes d'argon par une ouverture partielle ou totale du système : fusion partielle, métamorphisme, altération hydrothermale, réchauffement, déformation.

La méthode Ar-Ar permet de contourner en partie ce problème. Les différents ages obtenus par palier de températures correspondent à des dégazages d'argon situé dans des sites à énergie croissante d'activation. De ce fait, une faible crise métamorphique n'ayant affecté que les sites de basse énergie sera visualisée dans les tout premiers paliers, à basse température. Les sites à haute énergie d'activation, situés souvent au cœur des cristaux et restés pratiquement intact, ne libèreront l'argon qu'aux températures plus élevées, donnant ainsi un age plus important.

Problème de l'excès d'argon :

On parle d'excès d'argon quand l'âge obtenu est plus élevé que l'âge réel. Ce problème peut se produire en cas de présence d'argon hérité, non dégazé par l'évènement que l'on veut étudier, mais aussi dans le cas de diffusion d'argon étranger pouvant être capté dans les sites à basse énergie d'activation.

Signification et validité des ages plateaux pour le métamorphisme de faible degré :

On estime que pour avoir un âge cohérent avec l'age réel de l'épisode concerné, 4 paliers successifs doivent avoir le même age, appelé age plateau. La validité de datations de plans de foliations formés durant un métamorphisme de bas degré va donc être très dépendante de la nature polygénique des minéraux utilisés et de la quantité d'argon hérité. Dans le domaine de l'épimétamorphisme, ou proche de l'anchizone, la déformation n'est souvent pas assez pénétrative. Ainsi, les reliques des structures précédentes, ou la présence de minéraux détritiques ne permet pas une remise à zéro du chronomètre. On peut ainsi obtenir des spectres en « marche d'escalier », combinant les différents évènements tectonométamorphiques ayant subi l'échantillon. Par conséquent, une bonne sélection sur le terrain et une analyse pétrographique détaillée doit permettre d'exclure les échantillons avec des minéraux d'origines et d'âges différents. La qualité de la séparation joue aussi un grand rôle dans le sens ou elle va pouvoir permettre de sélectionner uniquement la fraction minérale que l'on veut dater. Cette étape permet parfois même de séparer les différentes générations d'une même espèce minérale, par exemple uniquement la fraction fine de la matrice.

Principes généraux de la méthode Ar-Ar

L'irradiation par neutrons rapides permet la formation d'un atome artificiel l'argon 39, à partir du potassium 39 naturel :

$$^{39}_{19}K(n,p)^{39}_{18}Ar$$

celui-ci est lui même radiaoctif, avec une période de 265ans :

$$^{39}Ar \xrightarrow{\beta^-} {}^{39}K$$

équations du rendement

$$^{39}Ar_s = {}^{39}K\Delta T \int_0^{+\infty} \Phi(E)\sigma(E)dE$$

soit

$$^{39}Ar_s = {}^{39}K\Delta T I$$

avec

$$I = \int_0^{+\infty} \Phi(E)\sigma(E)dE$$

Rappel sur le calcul de l'équation générale de la désintégration radioactive du système K-Ar :

$$^{40}Ar^* + {}^{40}Ca^* = {}^{40}K\left(e^{\lambda t} - 1\right)$$

avec

$$\lambda = \lambda_\varepsilon + \lambda_\beta$$

et

$$\frac{{}^{40}Ar^*}{{}^{40}Ca^*} = \frac{\lambda_\varepsilon}{\lambda_\beta}$$

on obtient alors

$${}^{40}Ar^* + {}^{40}Ar^* \frac{\lambda_\beta}{\lambda_\varepsilon} = {}^{40}K\left(e^{\lambda_t} - 1\right)$$

$${}^{40}Ar^*\left(1 + \frac{\lambda_\beta}{\lambda_\varepsilon}\right) = {}^{40}K\left(e^{\lambda_t} - 1\right)$$

$${}^{40}Ar^* = \frac{\lambda_\varepsilon}{\lambda} {}^{40}K\left(e^{\lambda_t} - 1\right)$$

soit

$$t_i = \frac{1}{\lambda}\ln\left[\frac{{}^{40}Ar^*}{{}^{40}K}\frac{\lambda}{\lambda_\varepsilon} + 1\right]$$

En combinant les équations de rendement et de désintégration radioactive du système K-Ar on obtient :

$$\frac{{}^{40}Ar^*}{{}^{39}Ar_s} = \frac{{}^{40}K\lambda_\varepsilon(e^{\lambda t_S} - 1)}{{}^{39}K\lambda\Delta TI}$$

Le paramètre J est tel que

$$J = \frac{{}^{39}K}{{}^{40}K}\frac{\lambda}{\lambda_\varepsilon}\Delta TI$$

soit

$$J = \frac{\left(e^{\lambda t_S} - 1\right)}{\left({}^{40}Ar^*/{}^{39}Ar\right)_S}$$

et enfin

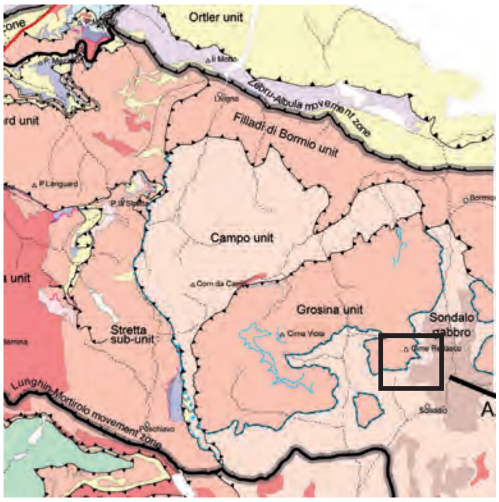
$$t_i = \frac{1}{\lambda}\ln\left[1 + \frac{\left({}^{40}Ar^*/{}^{39}Ar\right)_i}{\left({}^{40}Ar^*/{}^{39}Ar\right)_S}\left(e^{\lambda t_S} - 1\right)\right]$$

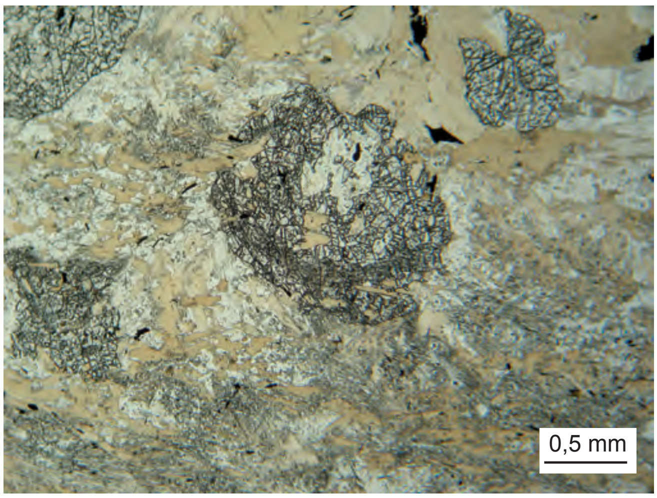
$$t_i = \frac{1}{\lambda}\ln\left[1 + J\left({}^{40}Ar^*/{}^{39}Ar\right)_i\right]$$

Sample 106

Amphibolite

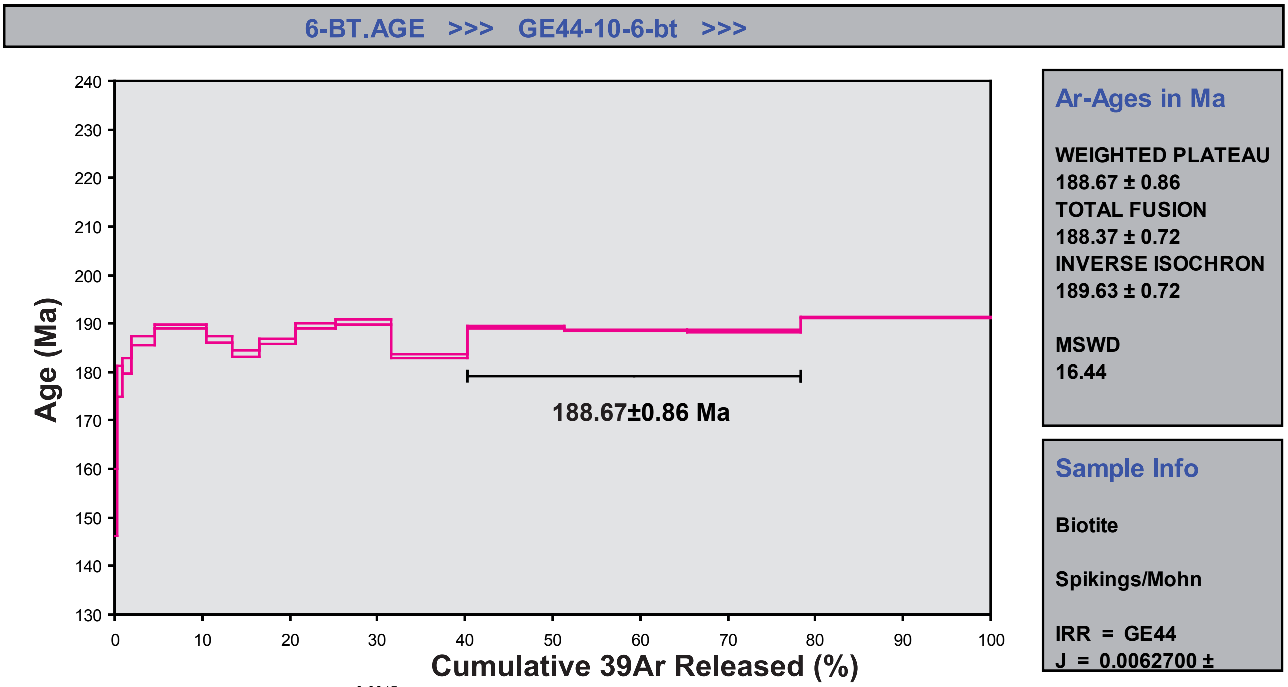
Coordinates: E 602403 N 5134738 Ref system: WGS 84, Coordinate system: UTM 32
Locality: Below Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on biotite
Tectonic setting: Sample located in the Campo basement in the vicinity (~30-40m) of the metmorphic aureole of the Sondalo gabbro
Sample description: Sample 106 is characterized by a compositional banding. Restite layer is made by sillimanite+biotite+quartz. Leucosome consists of biotite II+quartz+plagioclase. A younger generation of biotite III is locally observed in the cracks of the garnet.

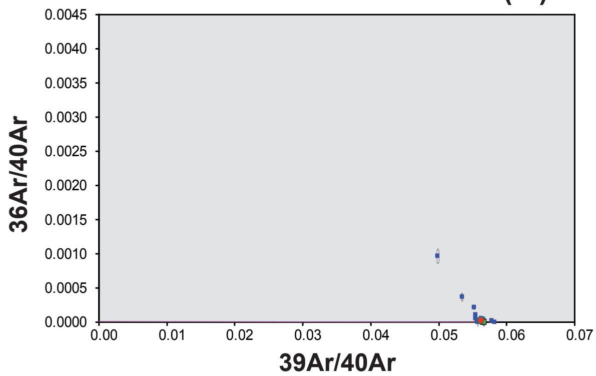




Sample 106

- Sample 106 contained three generations of biotite that could be defined on the basis of microstructural observations, which are probably responsible for the disturbed shape of the spectra.

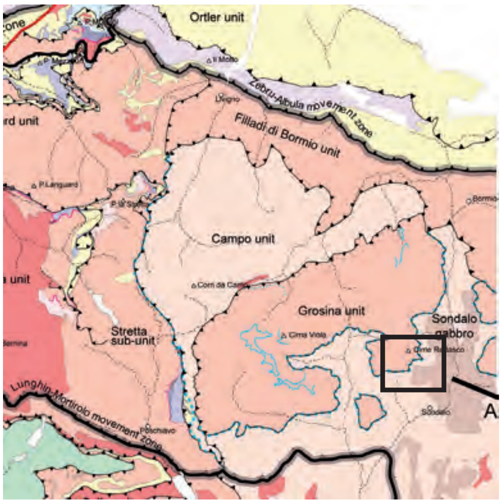


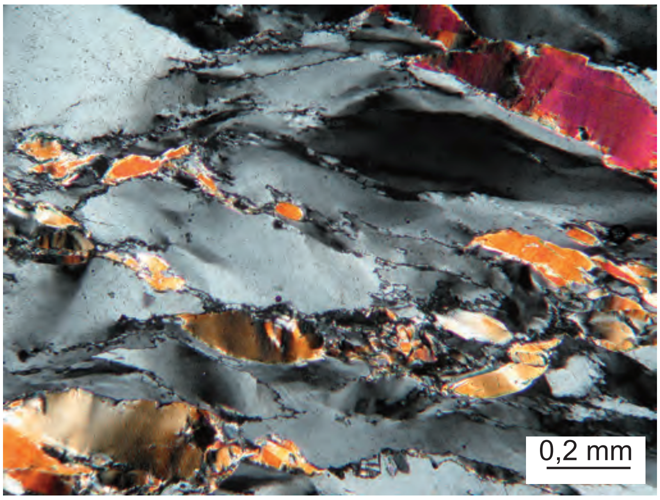


Laser (w)	40Ar/39 Ar	37Ar/39 Ar	36Ar/39 Ar	40Ar*/39A r_k	40Ar(mo l)	40Ar* (%)	39Ar $_k$ (%)	Age	±2 σ
6 biotite									
1,4	19,92750	0,00747	0,01966	14,117745	9,265E-16	70,85	0,28	153,02	6,92
1,7	18,60404	0,00903	0,00696	16,547378	1,707E-15	88,94	0,55	178,09	3,27
2,0	18,01731	0,00023	0,00394	16,852152	3,368E-15	93,53	1,11	181,21	1,63
2,2	17,95177	0,00010	0,00199	17,361803	7,996E-15	96,71	2,65	186,42	0,98
2,6	17,93330	0,00010	0,00093	17,658264	1,794E-14	98,47	5,96	189,44	0,33
2,9	17,66111	0,00010	0,00091	17,391271	8,382E-15	98,47	2,83	186,72	0,59
3,1	17,20395	0,00010	0,00036	17,097192	9,218E-15	99,38	3,19	183,72	0,63
3,5	17,56944	0,00010	0,00076	17,343906	1,192E-14	98,72	4,04	186,23	0,53
3,9	17,84099	0,00010	0,00059	17,666325	1,398E-14	99,02	4,67	189,52	0,52
4,3	17,85474	0,00010	0,00036	17,746593	1,898E-14	99,39	6,33	190,34	0,44
5,1	17,08047	0,00010	0,00008	17,054481	2,479E-14	99,85	8,64	183,28	0,31
6,0	17,67840	0,00015	0,00010	17,647339	3,272E-14	99,82	11,02	189,33	0,27
7,2	17,57790	0,00015	0,00015	17,576890	4,151E-14	99,99	14,06	188,61	0,19
8,5	17,55766	0,00015	0,00015	17,556655	3,831E-14	99,99	12,99	188,40	0,19
10,4	17,83295	0,00015	0,00015	17,831936	6,496E-14	99,99	21,69	191,21	0,18

Sample 117

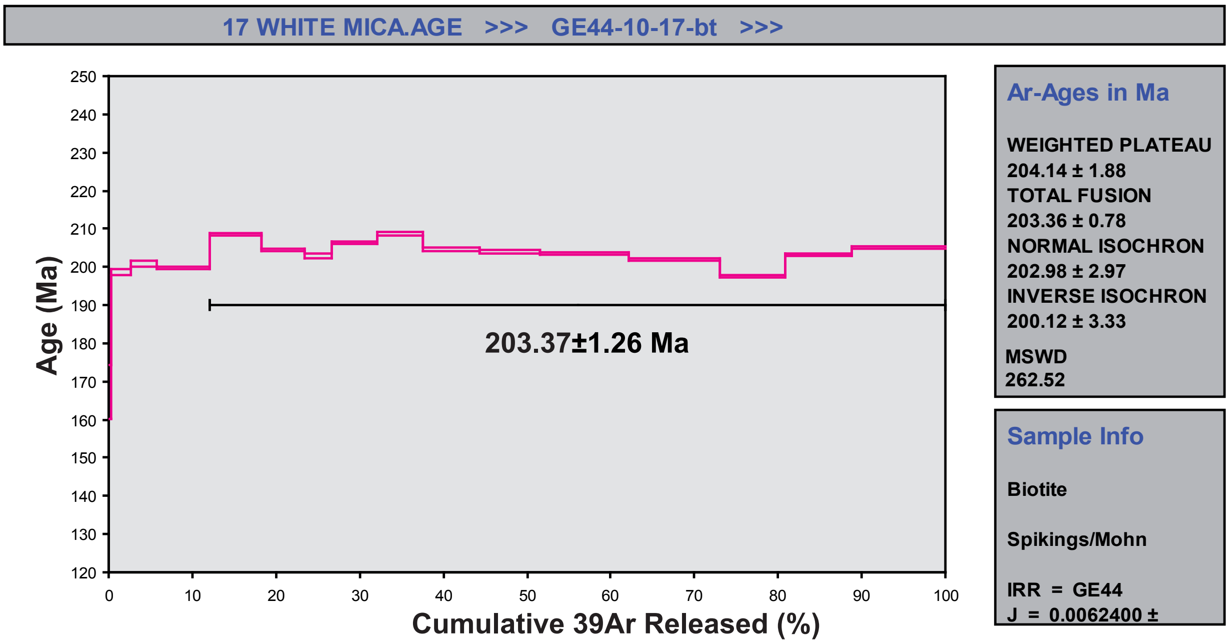
Muscovite
Coordinates: E 601888 N 5135219 Ref system: WGS 84, Coordinate system: UTM 32
Locality: Above Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on muscovite
Tectonic setting: Sample located in the Campo basement in the vicinity (~30-40m) of the metmorphic aureole of the Sondalo gabbro
Sample description: Sample 117 is defined by quartz+white mica +biotite locally retrogressed into chlorite

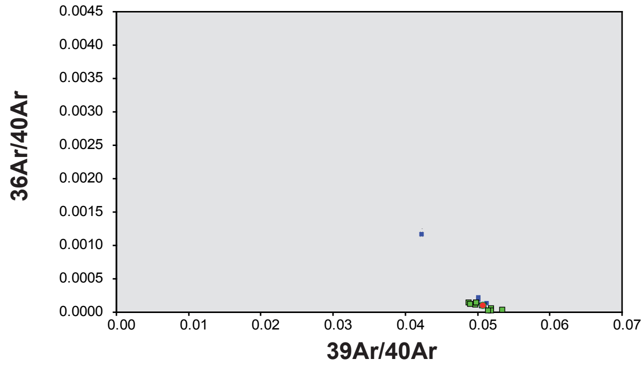




Sample 117

- Two generations of white mica are found in the sample, the rare fine grained aggregates of white mica have been selected out by hand picking. Therefore, the apparent age spectrum interpreted as resulting from the degassing of the argon reservoir related to the large porphyroclast white mica crystals

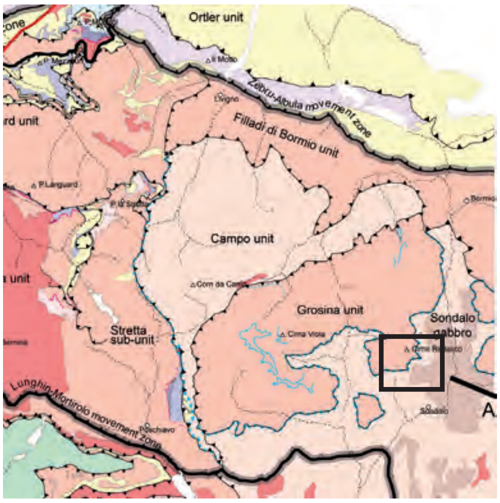


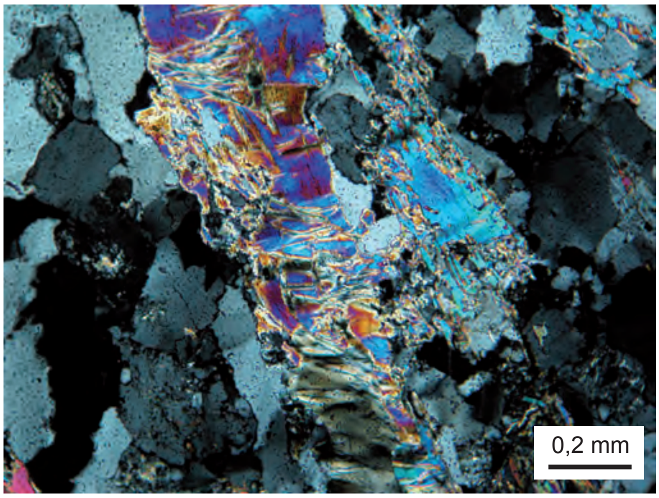


Laser (w)	40Ar/39Ar	37Ar/39Ar	36Ar/39Ar	$40Ar^*/39Ar_k$	40Ar(mol)	40Ar* (%)	$39Ar_k$ (%)	K/Ca	Age	±2σ
17 muscovite										
1,4	23,53434	0,00901	0,02792	15,283248	9,711E-16	64,94	0,38	47,718	165,04	7,35
1,7	19,81151	0,00064	0,00441	18,506146	6,116E-15	93,41	2,83	677,038	197,99	0,75
2,0	19,82142	0,00001	0,00366	18,738982	8,449E-15	94,54	3,91	1555,327	200,35	0,68
2,2	19,36670	0,00001	0,00249	18,630308	1,688E-14	96,20	7,99	1555,327	199,25	0,39
2,7	19,93616	0,00001	0,00285	19,093693	1,382E-14	95,77	6,35	1555,327	203,94	0,42
3,1	19,53512	0,00001	0,00199	18,945903	8,582E-15	96,98	4,03	1555,327	202,44	0,55
3,4	19,97448	0,00001	0,00231	19,289609	1,475E-14	96,57	6,77	1555,327	205,91	0,39
4,2	19,92824	0,00001	0,00278	19,104887	1,818E-14	95,87	8,36	1555,327	204,05	0,42
4,8	19,93046	0,00001	0,00296	19,053345	1,950E-14	95,60	8,97	1555,327	203,53	0,41
5,6	19,60944	0,00001	0,00207	18,996279	2,820E-14	96,87	13,18	1555,327	202,95	0,31
6,5	19,16455	0,00001	0,00105	18,852912	2,825E-14	98,37	13,51	1555,327	201,50	0,35
8,8	19,14233	0,00001	0,00049	18,997759	2,069E-14	99,24	9,91	1555,327	202,97	0,35
10,4	19,29823	0,00001	0,00038	19,186045	2,914E-14	99,42	13,84	1555,327	204,87	0,29

Sample 19

Migmatite
Coordinates: E 601861 N 5134869 Ref system: WGS 84, Coordinate system: UTM 32
Locality: On the path, above Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on muscovite
Tectonic setting: Sample located in the Campo basement in the metmorphic aureole (~30-40m from the contact) of the Sondalo gabbro
Sample description: Quartz+biotite+plagioclase+fine-grained muscovite II and porphyroclast of muscovite I

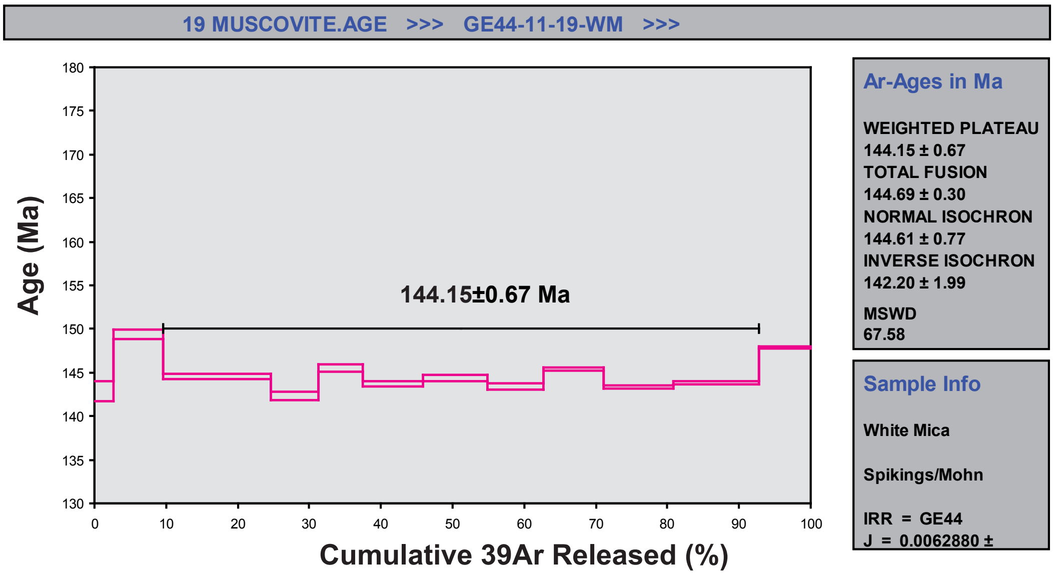


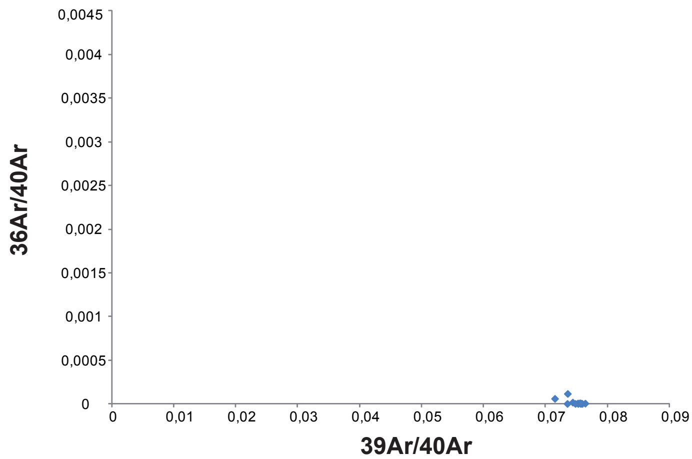


Sample 19

- Sample 19 contained two generations of white micas that could be defined on the basis of microstructural data giving probably the disturbing shape of the spectra.

- We can propose that the spectrum is the result of the mixing between pre-Permian or Permian white micas with a younger generation (Jurassic or Alpine age).



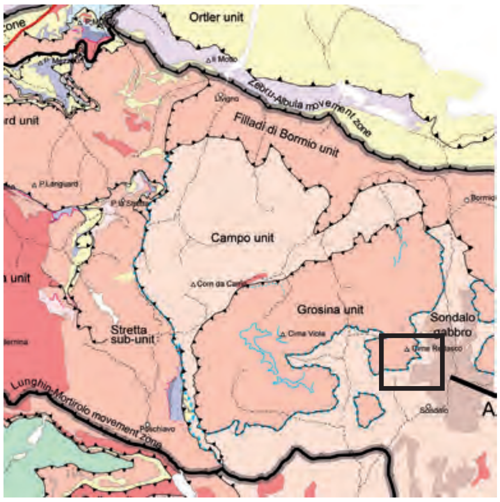


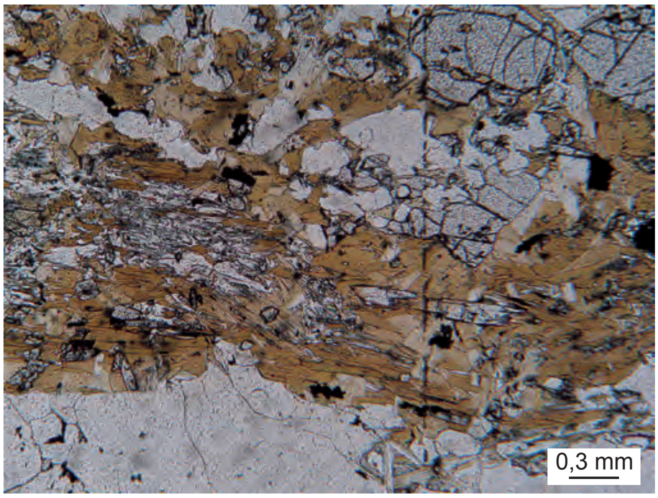
Laser (w)	40Ar/39 Ar	37Ar/39 Ar	36Ar/39 Ar	40Ar*/39A r_k	40Ar (mol)	40Ar* (%)	$39Ar_k$ (%)	Age	±2σ
19 muscovite									
J=0.0062880±0.0000125									
1,7	13,57622	0,00001	0,00158	13,108761	3,711E-15	96,56	2,66	142,89	1,16
2,0	13,97059	0,00001	0,00083	13,725621	9,935E-15	98,25	6,91	149,35	0,53
2,2	13,26827	0,00001	0,00001	13,265451	2,066E-14	99,98	15,13	144,54	0,29
2,7	13,06958	0,00001	0,00006	13,052193	8,798E-15	99,87	6,54	142,30	0,52
3,1	13,42683	0,00001	0,00022	13,361320	8,664E-15	99,51	6,27	145,54	0,45
3,4	13,21738	0,00001	0,00011	13,183982	1,138E-14	99,75	8,37	143,68	0,33
4,2	13,27663	0,00001	0,00008	13,252482	1,223E-14	99,82	8,95	144,40	0,38
4,8	13,17605	0,00002	0,00006	13,158490	1,056E-14	99,87	7,79	143,42	0,38
5,6	13,34992	0,00002	0,00001	13,348914	1,161E-14	99,99	8,45	145,41	0,15
6,5	13,15489	0,00002	0,00001	13,153876	1,314E-14	99,99	9,71	143,37	0,14
8,8	13,19845	0,00001	0,00001	13,197435	1,627E-14	99,99	11,99	143,82	0,14
10,4	13,58556	0,00001	0,00001	13,584554	1,009E-14	99,99	7,22	147,87	0,16

Sample 21

Migmatite

Coordinates: E 602443 N 5134718 Ref system: WGS 84, Coordinate system: UTM 32
Locality: Close to path, below Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on biotite
Tectonic setting: Sample located in the Campo basement in the vicinity (~30-40m) of the metmorphic aureole of the Sondalo gabbro
Sample description: Sample 21 is characterized by a compositional banding. Restite layer is made by sillimanite+biotite+quartz. Leucosome is made by biotite II+quartz+plagioclase, locally a younger generation of biotite III is developing in the cracks of the garnet.

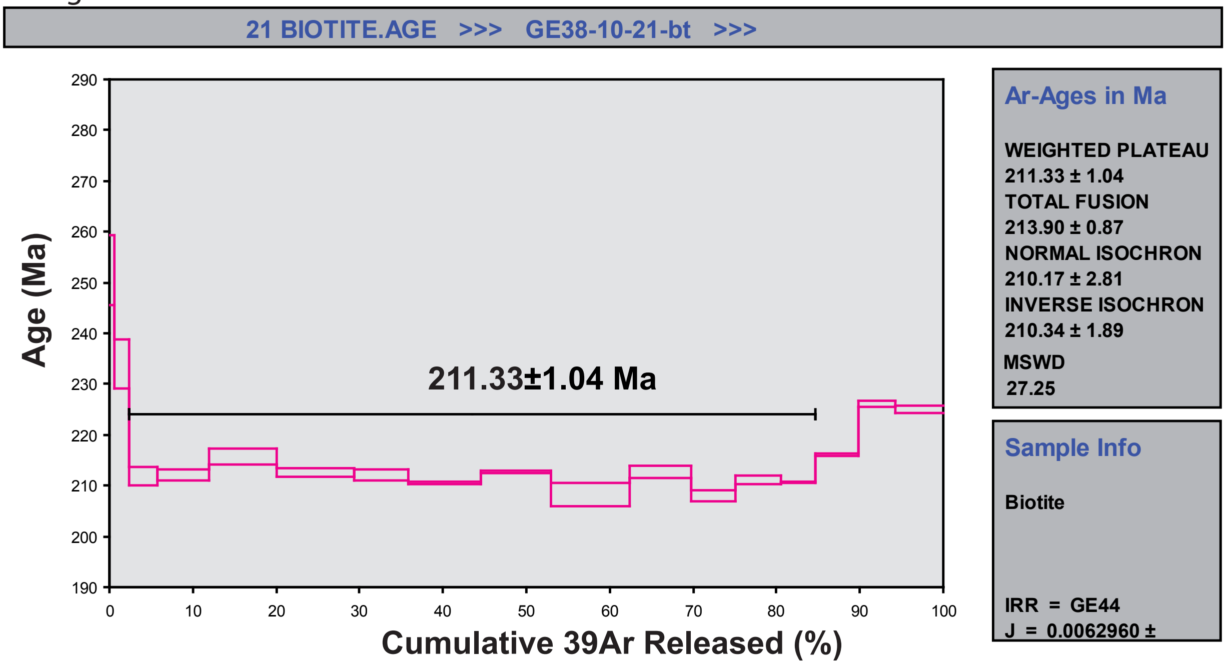


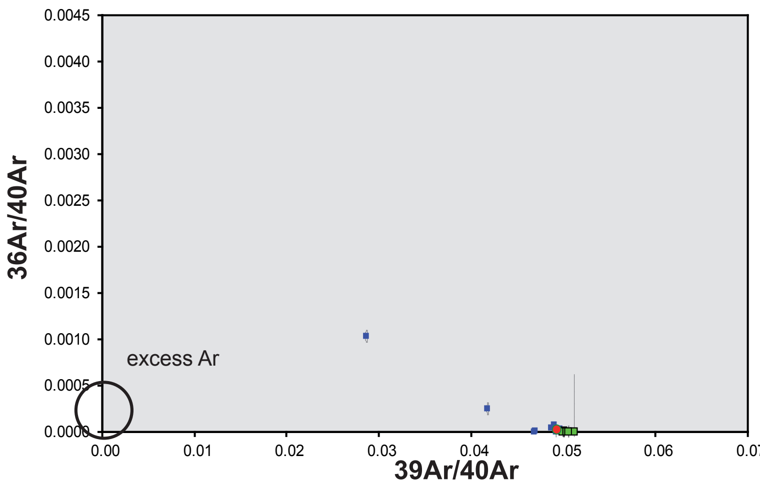


Sample 21

- Sample 21 contained three generations of biotite that could be defined on the basis of microstructural data giving probably the disturbing shape of the spectra.

- Data points have a tendency to disperse, which could be a consequence of the relative contributi "excess argon"

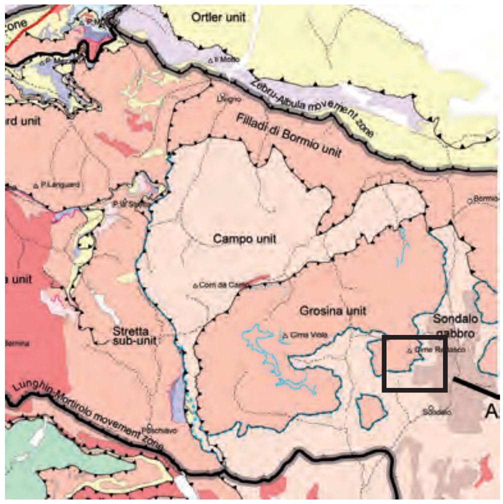


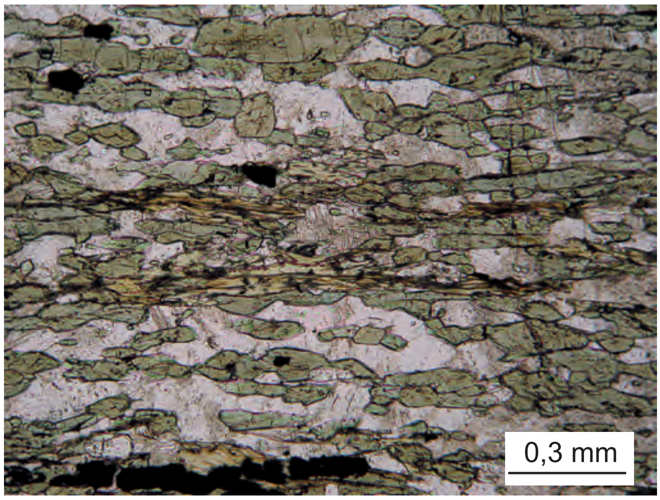


Laser (w)	40Ar/39Ar	37Ar/39Ar	36Ar/39Ar	$40Ar^*/39Ar_k$	40Ar(mol)	40Ar* (%)	$39Ar_k$ (%)	K/Ca	Age
21 biotite									
1,4	34,68117	0,00747	0,03663	23,855260	1,380E-15	68,78	0,53	57,548	252,4
1,7	23,77535	0,00001	0,00606	21,983132	3,362E-15	92,46	1,89		233,8
2,0	20,27576	0,00001	0,00162	19,796975	5,104E-15	97,64	3,37		211,9
2,2	20,02480	0,00001	0,00072	19,810401	9,245E-15	98,93	6,19		212,0
2,6	20,41716	0,00288	0,00083	20,171684	1,230E-14	98,80	8,07	149,073	215,7
2,9	19,90063	0,00001	0,00011	19,868263	1,385E-14	99,84	9,32	217,845	212,6
3,1	19,87165	0,00001	0,00016	19,823359	9,609E-15	99,76	6,48	217,845	212,1
3,4	19,66735	0,00001	0,00000	19,666343	1,280E-14	99,99	8,72	217,845	210,6
3,8	19,87453	0,00001	0,00002	19,873522	1,252E-14	99,99	8,44	217,845	212,6
4,2	19,43516	0,00001	0,00002	19,434146	1,364E-14	99,99	9,40	217,845	208,2
4,6	19,90226	0,00091	0,00009	19,874200	1,080E-14	99,86	7,27	472,167	212,6
4,9	19,43370	0,00002	0,00005	19,416791	7,694E-15	99,91	5,30	1158,878	208,0

Sample 31

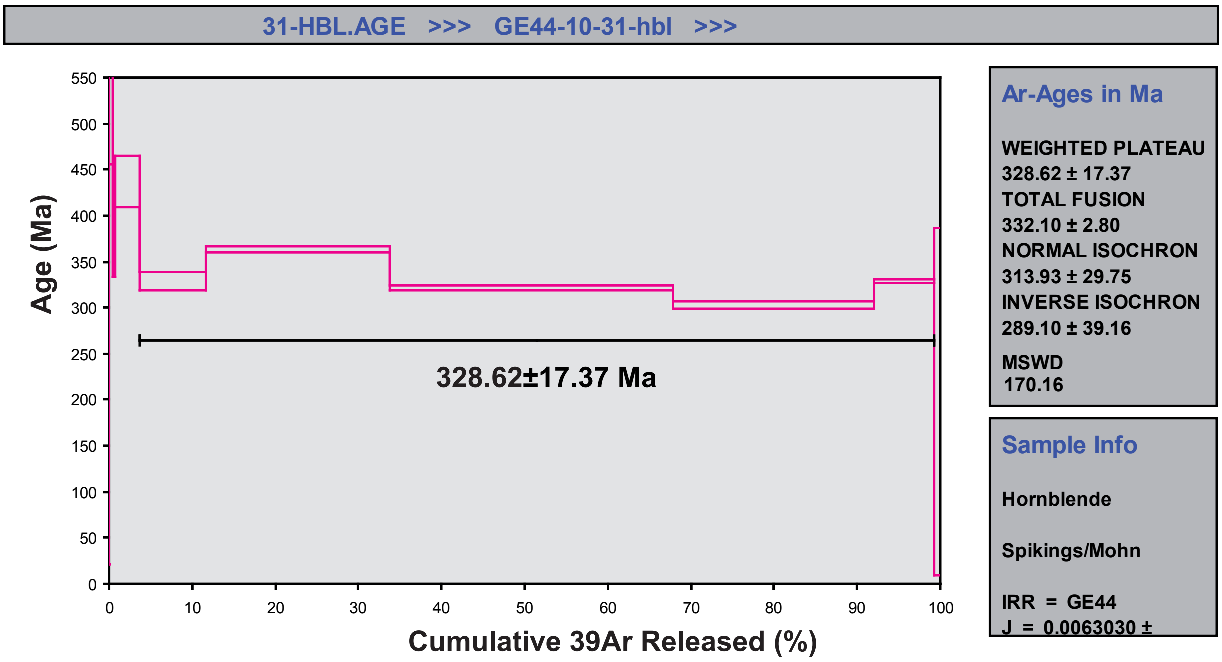
Amphibolite
Coordinates: E 601943 N 5135113 Ref system: WGS 84, Coordinate system: UTM 32
Locality: Above Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on amphibole
Tectonic setting: Sample located in the Campo basement in the vicinity (~30-40m) of the metmorphic aureole of the Sondalo gabbro
Sample description: amphibole+epidote+chlorite+plagioclase

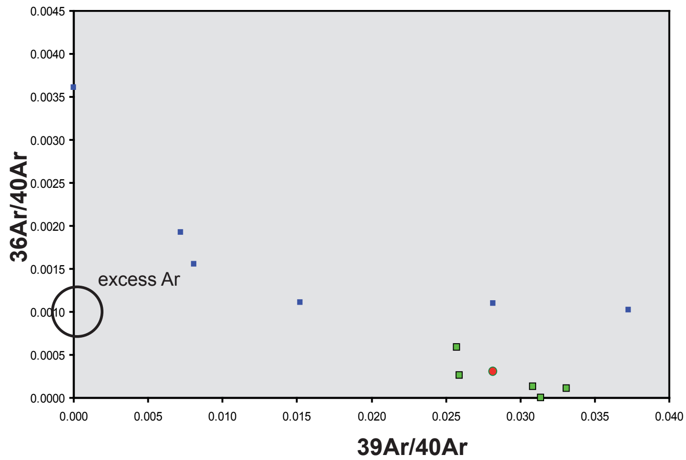


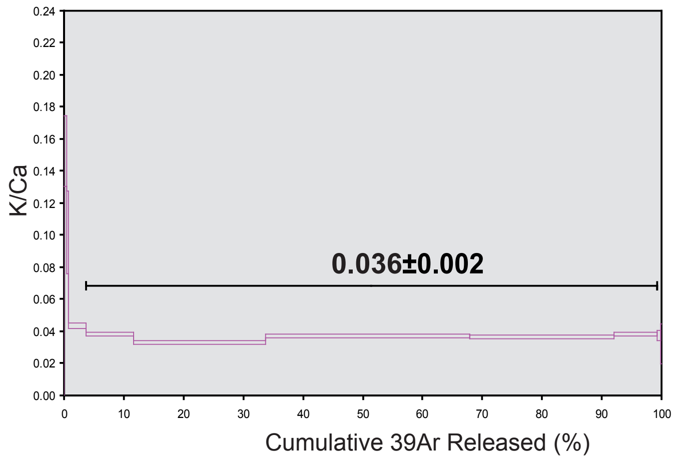


Sample 31

- One generation of amphibole in equilibrium with minerals typical of greenschist facies condition.
- Data points have a tendency to disperse, which could be a consequence of the relative contribution of the "excess argon"



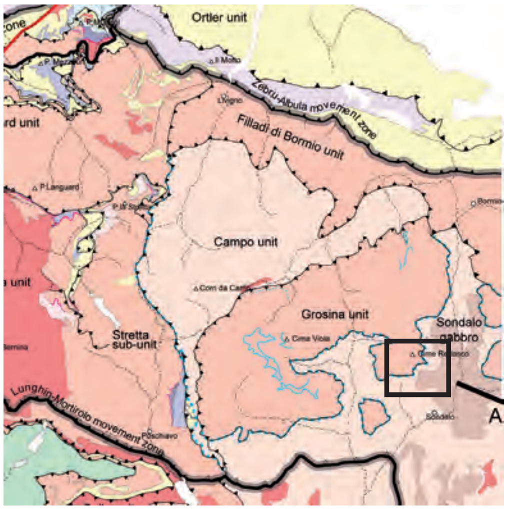


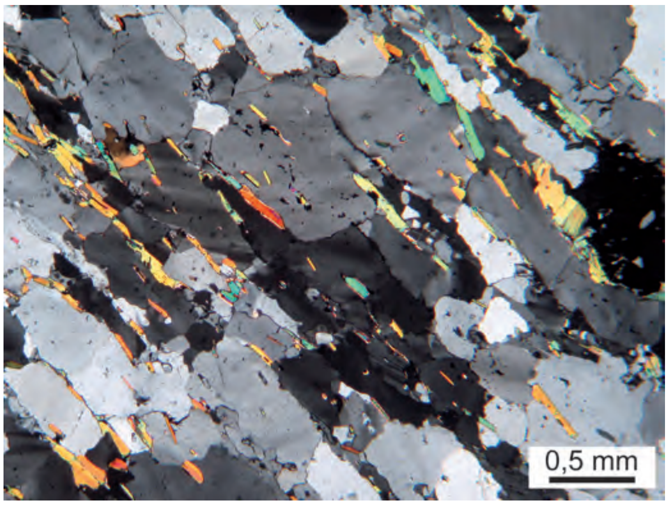


Laser	40Ar/39Ar	37Ar/39Ar	36Ar/39Ar	$40Ar^*/39Ar_k$	40Ar(mol)	40Ar* (%)	$39Ar_k$ (%)	K/Ca	Age	±2σ
Power (w)										
31 hornblende										
1,4	0,00001	0,00001	0,00001	1,893204	1,302E-17	0,00	0,00	0,000	21,40	0,00
1,7	138,17861	2,84969	0,27242	58,011891	2,288E-16	41,90	0,49	0,151	562,20	106,50
2,0	122,18002	4,26804	0,19543	64,949997	1,172E-16	53,01	0,28	0,100	619,07	285,49
2,3	64,96102	9,91613	0,07626	43,490316	6,586E-16	66,50	2,95	0,043	437,05	27,47
2,7	38,31641	11,35009	0,02606	31,743983	1,044E-15	82,21	7,93	0,038	329,05	10,14
3,1	38,01992	13,04443	0,01351	35,355079	2,883E-15	92,17	22,05	0,033	362,94	3,20
3,9	31,99495	11,60768	0,00744	30,942113	3,754E-15	95,95	34,16	0,037	321,43	2,28
4,7	29,76369	11,87490	0,00654	28,988037	2,477E-15	96,62	24,23	0,036	302,74	3,79
5,5	31,44248	11,35436	0,00001	31,683590	7,702E-16	100,00	7,13	0,038	328,47	2,26
6,2	26,46654	11,59642	0,03081	18,409145	5,755E-17	69,01	0,63	0,037	198,05	188,71
7,3	35,00362	13,43726	0,04310	23,528071	1,701E-17	66,61	0,14	0,032	249,47	648,77

Sample 132

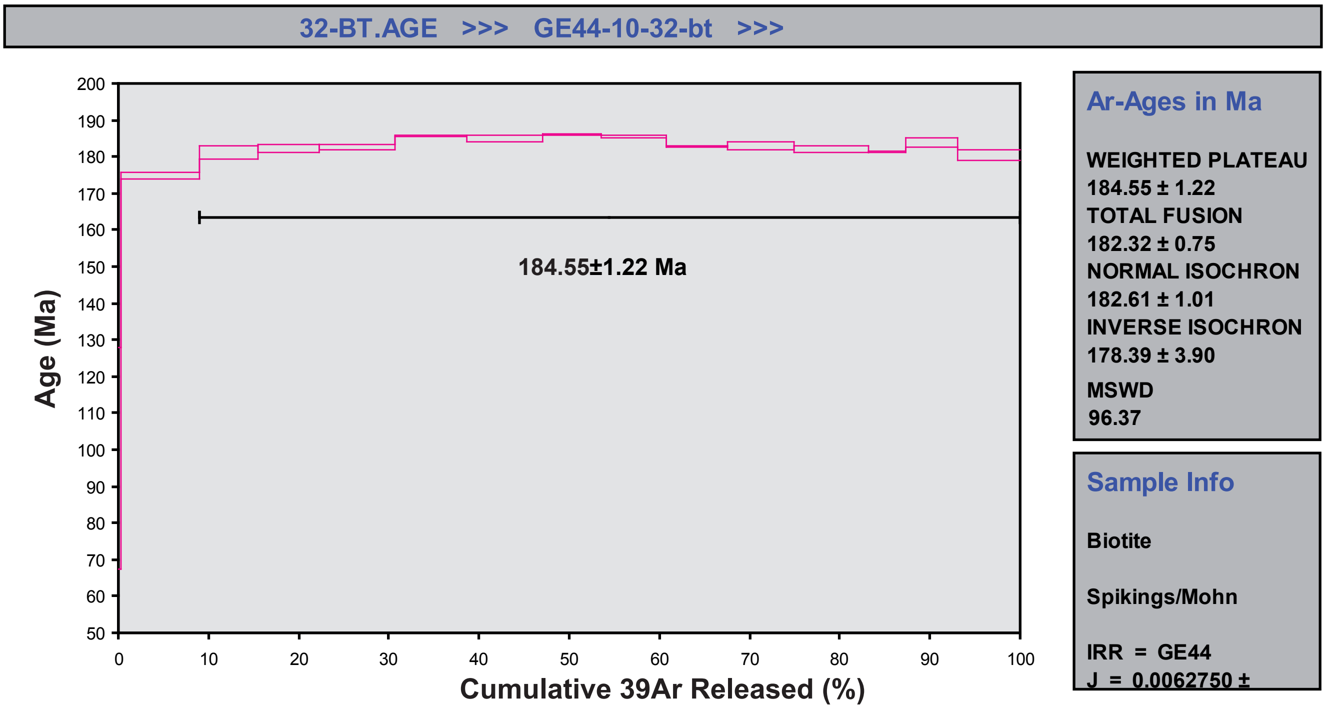
Amphibolite
Coordinates: E 601932 N 5135182 Ref system: WGS 84, Coordinate system: UTM 32
Locality: Above Alto, near Sondalo, Valtellina valley
Dating: Ar/Ar dating on biotite
Tectonic setting: Sample located in the Campo basement in the vicinity (~30-40m) of the metmorphic aureole of the Sondalo gabbro
Sample description: Pervasive foliation defined by the preferred orientation of white mica+biotite in equilibrium with quartz+garnet+plagioclase

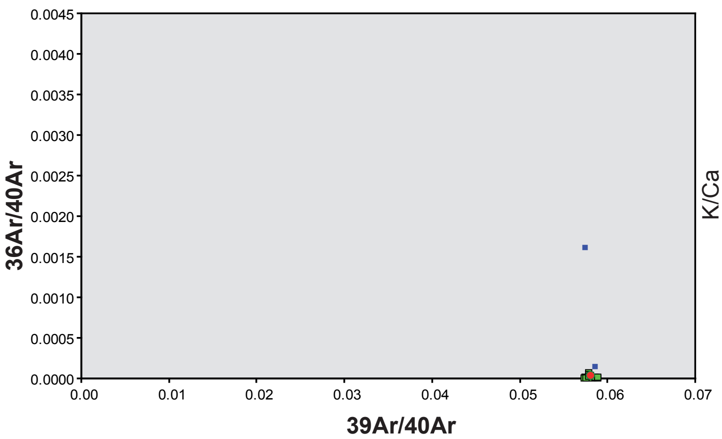


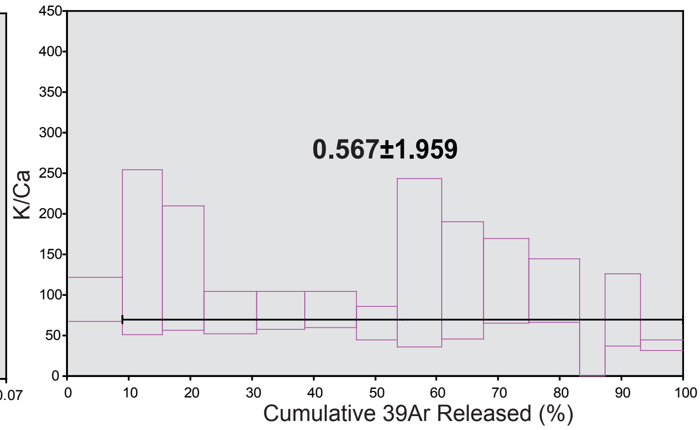


Sample 132

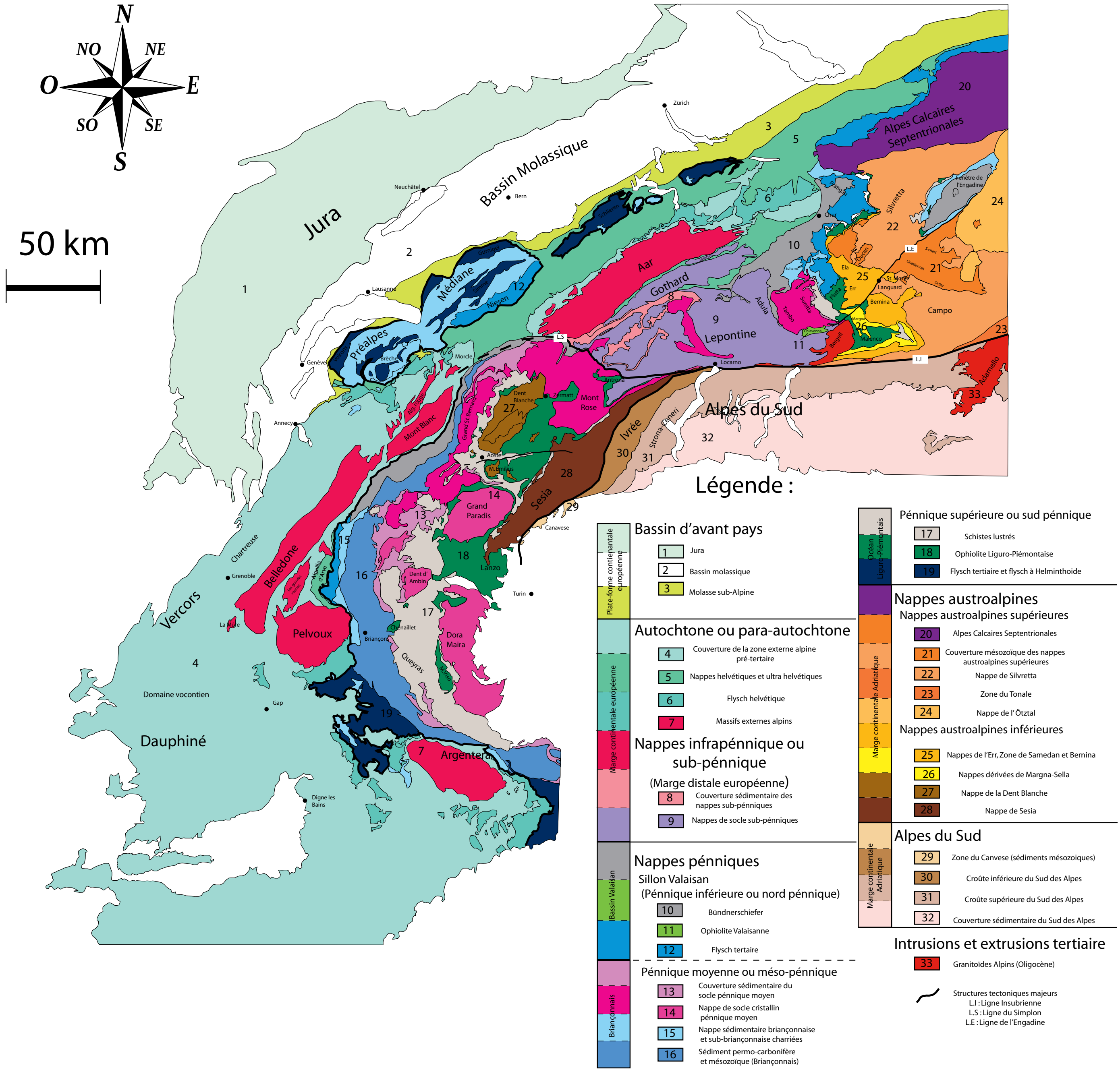
- One generation of biotite which is stable with garnet and plagioclase in the Permian fabric related to the Sondalo intrusion.



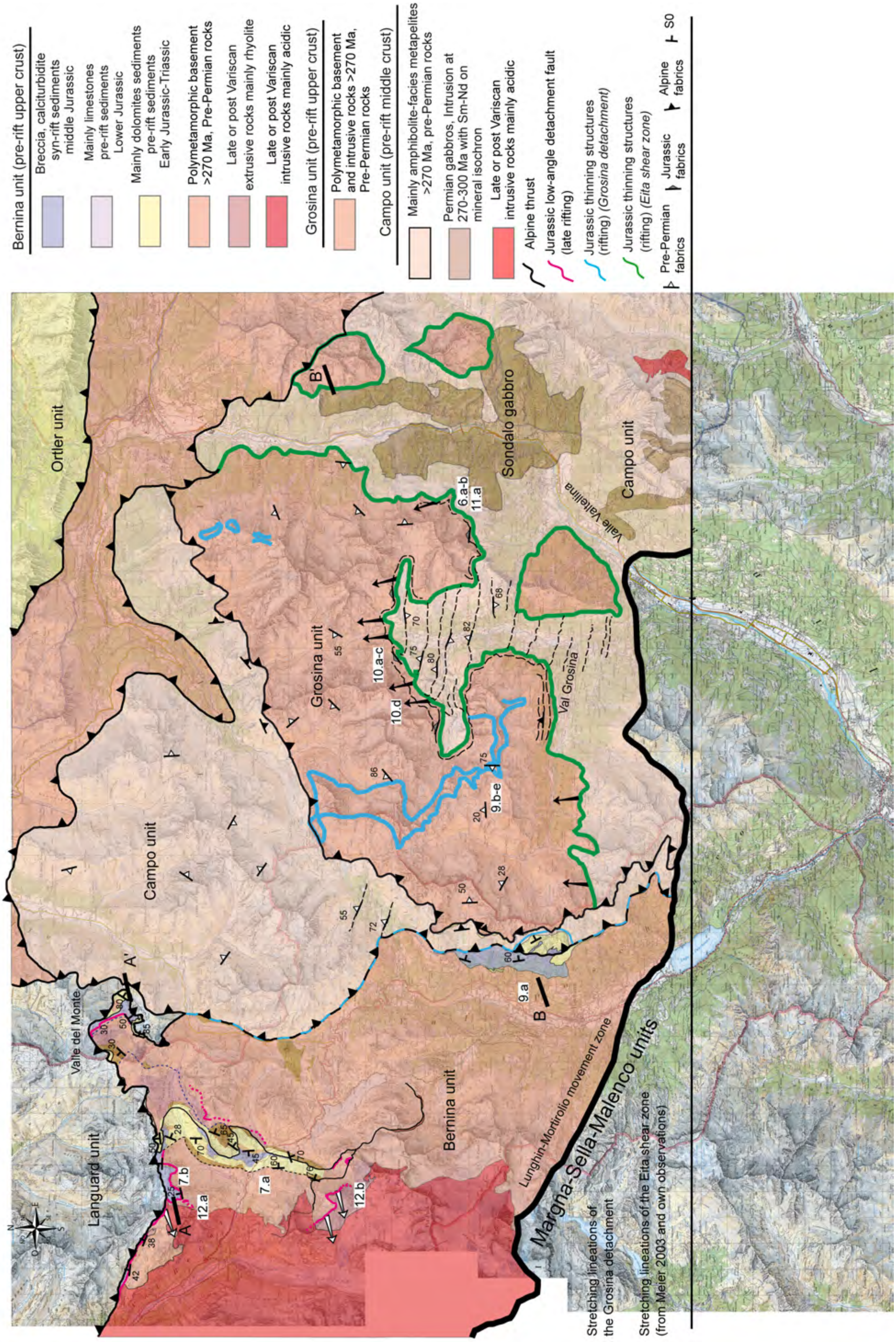




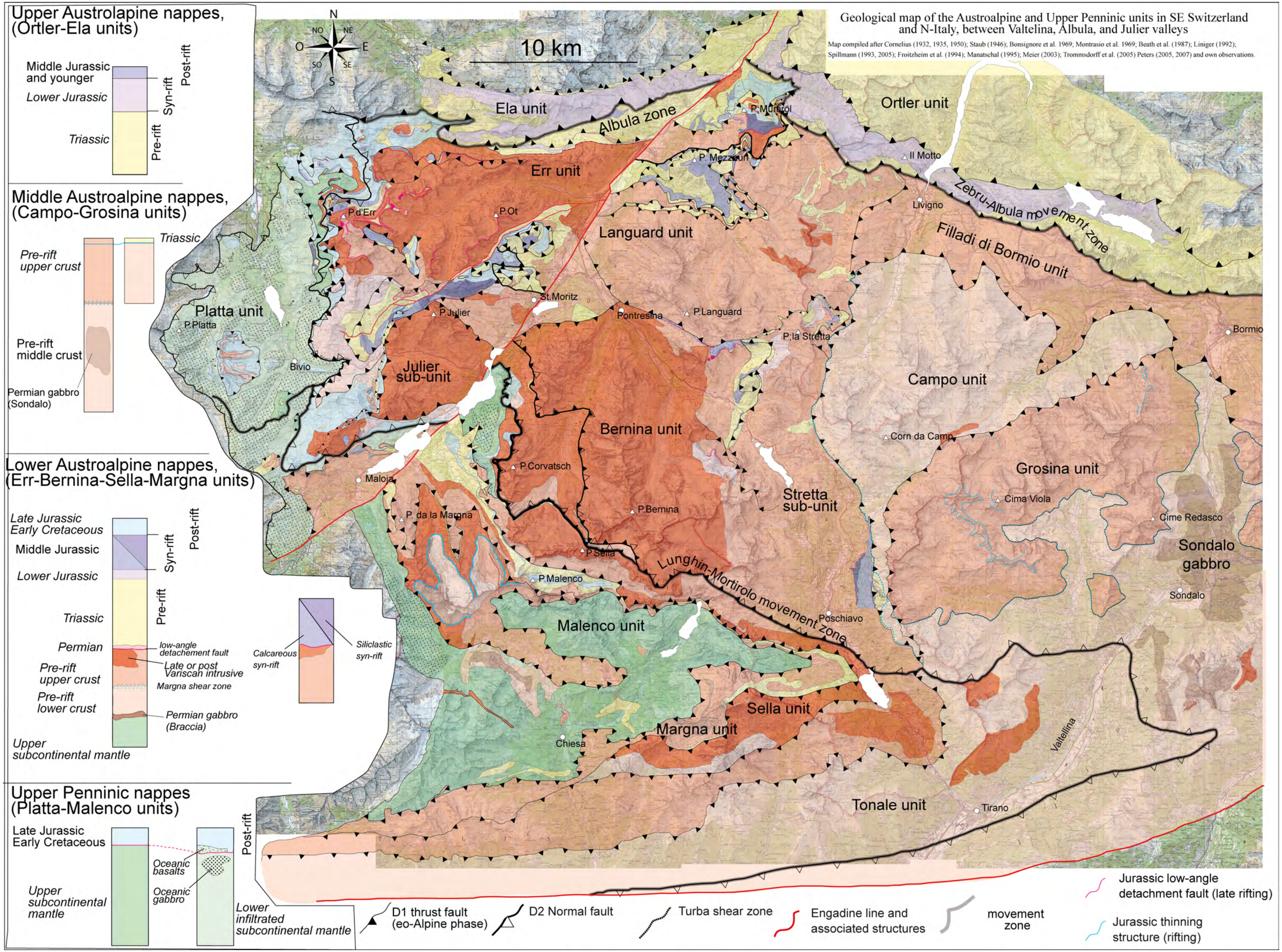
Laser (w)	40Ar/39Ar	37Ar/39Ar	36Ar/39Ar	$40Ar^*/39Ar_k$	40Ar(mol)	40Ar* (%)	$39Ar_k$ (%)	K/Ca	Age	±2σ
32 biotite										
1,4	17,28995	0,00002	0,02851	8,863757	1,837E-16	51,27	0,31	100,769	97,65	30,26
1,7	16,96400	0,00002	0,00251	16,221879	5,066E-15	95,63	8,66	93,585	174,88	0,93
2,0	17,15386	0,00285	0,00108	16,834160	3,878E-15	98,14	6,55	150,914	181,16	1,77
2,2	17,18308	0,00327	0,00082	16,938737	4,007E-15	98,58	6,76	131,410	182,23	1,19
2,6	17,12051	0,00558	0,00050	16,973188	4,999E-15	99,14	8,46	77,111	182,59	0,86
2,9	17,26758	0,00537	0,00001	17,266628	4,713E-15	99,99	7,91	80,103	185,59	0,21
3,1	17,29408	0,00533	0,00026	17,216628	4,984E-15	99,55	8,35	80,641	185,08	0,92
3,5	17,32040	0,00665	0,00002	17,319463	3,902E-15	99,99	6,53	64,652	186,13	0,21
3,9	17,26482	0,00312	0,00002	17,263845	4,333E-15	99,99	7,27	137,886	185,56	0,25
4,3	16,99212	0,00369	0,00001	16,991153	3,980E-15	99,99	6,79	116,468	182,77	0,25
5,1	17,04185	0,00370	0,00009	17,013126	4,328E-15	99,83	7,36	116,103	183,00	1,22
6,0	16,95342	0,00414	0,00010	16,921916	4,836E-15	99,81	8,27	103,847	182,06	1,01
7,2	16,85759	0,00001	0,00001	16,856582	2,372E-15	99,99	4,08		181,39	0,30
8,5	17,17520	0,00535	0,00030	17,086371	3,397E-15	99,48	5,73	80,447	183,74	1,25
10,4	16,85288	0,01143	0,00027	16,771988	4,044E-15	99,52	6,96	37,614	180,53	1,43



Carte tectonique des Alpes (modifié d’après Schmid et al. 2004)



Carte géologique de la region entre Val da Fain et Val Poschiavo



Geological map of the Austroalpine and Upper Penninic units in SE Switzerland and N-Italy, between Valtelina, Albula, and Julier valleys
Map compiled after Cornelius (1932, 1935, 1950); Staub (1946); Bonsignore et al. 1969; Montrasio et al. 1969; Beath et al. (1987); Liniger (1992); Spillmann (1993, 2005); Froitzheim et al. (1994); Manatschal (1995); Meier (2003); Trommsdorff et al. (2005) Peters (2005, 2007) and own observations.
10 km
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NO
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Upper Austroalpine nappes, (Ortler-Ela units)
Middle Jurassic and younger
Lower Jurassic
Triassic
Post-rift
Syn-rift
Pre-rift
Middle Austroalpine nappes, (Campo-Grosina units)
Triassic
Pre-rift upper crust
Pre-rift middle crust
Permian gabbro (Sondalo)
Lower Austroalpine nappes, (Err-Bernina-Sella-Margna units)
Late Jurassic Early Cretaceous
Middle Jurassic
Lower Jurassic
Triassic
Permian
low-angle detachement fault
Late or post Variscan intrusive
Pre-rift upper crust
Margna shear zone
Pre-rift lower crust
Permian gabbro (Braccia)
Upper subcontinental mantle
Post-rift
Syn-rift
Pre-rift
Calcareous syn-rift
Siliclastic syn-rift
Upper Penninic nappes (Platta-Malenco units)
Late Jurassic Early Cretaceous
Oceanic basalts
Oceanic gabbro
Upper subcontinental mantle
Lower infiltrated subcontinental mantle
Post-rift
Ela unit
Albula zone
P.Murtirol
Ortler unit
Il Motto
P. Mezzaun
Err unit
P.d'Err
P.Ot
Livigno
Zebru-Albula movement zone
Filladi di Bormio unit
Languard unit
Platta unit
P.Platta
St.Moritz
P.Julier
Pontresina
P.Languard
P.la Stretta
Bormio
Julier sub-unit
Bivio
Campo unit
Bernina unit
Corn da Camp
Grosina unit
Maloja
P.Corvatsch
Stretta sub-unit
Cima Viola
Cime Redasco
P. da la Margna
P.Bernina
Sondalo gabbro
P.Sella
Lunghin-Mortirolo movement zone
P.Malenco
Sondalo
Malenco unit
Poschiavo
Sella unit
Margna unit
Chiesa
Valtellina
Tonale unit
Tirano
D1 thrust fault (eo-Alpine phase)
D2 Normal fault
Turba shear zone
Engadine line and associated structures
movement zone
Jurassic low-angle detachment fault (late rifting)
Jurassic thinning structure (rifting)