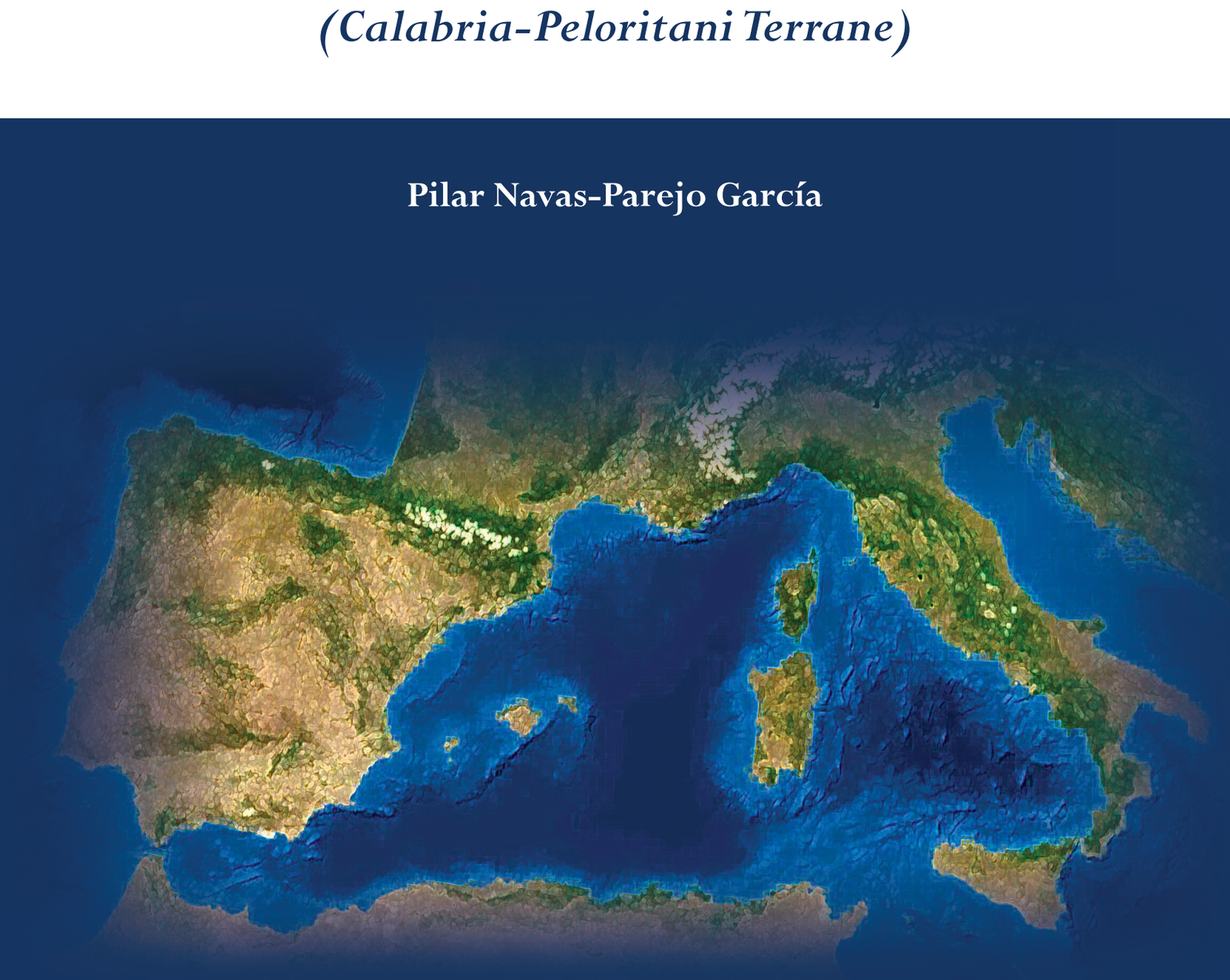
Paleozoic Stratigraphy and Palaeogeography of the Malaguide Complex (Betic Cordillera) and other Western Mediterranean related domains (Calabria-Peloritani Terrane)

Pilar Navas-Parejo García



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Paleozoic Stratigraphy and Palaeogeography of the Malaguide Complex (Betic Cordillera) and other Western Mediterranean related domains (Calabria-Peloritani Terrane)

Estratigrafía y Paleogeografía del Complejo Maláguide (Cordillera Bética) y otros dominios relacionados del Mediterráneo occidental (Terreno Calabro-Peloritano)

Memoria de Tesis presentada por Dña. Pilar Navas-Parejo García para optar al grado de Doctora con mención internacional por la Universidad de Granada.

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A mi tía, la primera Pilar Navas-Parejo

Ex-libro lapidum historia mundi

Abstract

This thesis studies the Paleozoic stratigraphy, palaeogeography and geo-historical evolution of the intra-Alpine terrains outcropping in the Betic Cordillera and in the equivalent Alpine domains of the Calabria-Peloritani Terrane. Paleozoic deposits studied in the Betic Cordillera belong to the Malaguide Complex, the upper tectonic complex in the Internal Domains. In Calabria, the tectonic unit whose Paleozoic terrains show more resemblance with those of the Malaguide Complex is the Stilo Unit, which crops out in the Serre and Aspromonte massifs and which also occurs the highest tectonic position within the Internal Domains stack. Some Malaguide affinities, although also important differences, are stated in the Longi-Taormina Unit, which only crops out in the Sicilian Peloritani Mountains, at the base and to the front of the Internal Domains stack along the tectonic contact with more external domains, then in a different tectonic position compared to those of the Stilo Unit and of the Malaguide Complex.

All successions here studied are marine, relatively deep, predominantly terrigenous-clastic turbiditic, lithologically monotonous, poor in fossils, strongly deformed during Alpine and Variscan orogenies, and frequently affected by a low- or very low-grade metamorphism. All those facts make lithostratigraphic studies, dating and correlation, as well as a correct interpretation of the palaeogeographic significance remarkably difficult. However, these successions contain thin and scarce carbonatic and siliceous (radiolaritic) intercalations that, in some cases, yielded fossils, conodonts in particular, which allow detailed stratigraphic studies and reliable correlations. The present Doctoral thesis has been focused precisely in the lithostratigraphic description, biostratigraphic dating by means of conodonts, correlation and comparative analysis of several selected successions in the tectonic units above mentioned. The main aim has been precising the fundamental features of their palaeogeography and geo-historical Paleozoic evolution, which are still badly constrained. For that purpose 18 sections in the Malaguide Complex, 8 sections in the Stilo Unit and 6 sections in the Longi-Taormina Unit have been selected for their litho- and biostratigraphic study. Special attention has been paid to dating of the carbonatic horizons by means of conodonts, which is the more effective fossil group to cover the stratigraphic and palaeogeographic aims proposed. The studied areas in the Malaguide Complex are located in the provinces of Almería and Murcia (Vélez Rubio-Lorca Corridor), Granada (Cogollos Vega Zone) and Málaga (Ardales area). In southern Italy, the Stilo Unit has been studied in two sectors, located in the Stilo-Bivongi region (Serre) and in the Stati-Pietrapennata area (S Aspromonte). The Longi-Taormina Unit, which crops out exclusively in northern Sicily (Peloritani Mountains), has been mainly studied in their central and eastern outcrops.

Field, stratigraphic and biostratigraphic studies have allowed the division of the Malaguide Complex outcropping in all the three studied areas in three Alpine tectonic units, which are, from bottom to top, Epimetamorphic Malaguide Unit, Lower Malaguide Unit and Upper Malaguide Unit. These units show, in addition, Paleozoic successions quite different among them. In these units, the Upper Ordovician (Hirnantian) in the Ardales region, the Devonian (Emsian and Famennian), and the Lower Carboniferous (Visean) in the Cogollos Vega Zone, and the Upper Carboniferous (Bashkirian) in the Vélez Rubio-Lorca Corridor have been dated, for the first time in this study, by

means of conodonts.

The lithostratigraphic succession in the Epimetamorphic Malaguide Unit is clearly of Malaguide-type and similar to that of the Lower Malaguide Unit. The metamorphic imprint that affects it and its tectonic position below more typical Malaguide terrains of the Lower Malaguide Unit, have led, however, to some authors to includes part of its outcrops in the Alpujarride Complex. The Lower Malaguide Unit includes the classical, more thick and widespread pre-Carboniferous Malaguide succession. This succession is constituted by the Morales and Santi Petri formations, the later characterized by the classical calizas alabeadas facies. The pre-Carboniferous succession of the Upper Malaguide Unit is less thick and more complicated in detail, from both a tectonic and a stratigraphic point of view. The Upper Malaguide pre-Carboniferous succession is characterized by the presence of thin stratigraphic horizons of limestones bearing Silurian-Devonian conodonts, which are laterally discontinuous and are intercalated within pelites; it is the only Malaguide unit in the studied areas that includes the Marbella Formation. On the contrary, the Lower Carboniferous formations are equivalent in all the Malaguide units and are characterized by the Falcoña Formation (very thin, that includes Tournaisian lydites -radiolarites- and Visean conodont-bearing limestones) and the Almogía Formation (very thick, Serpukhovian turbiditic siciliclastic Culm facies).

Silurian-Devonian conodont-bearing limestone horizons also appear, exceptionally, in the Epimetamorphic Malaguide Unit (González sections) and in the Lower Malaguide Unit (Peña del Castillo section). In the Ardales area, the subdivision of the Lower Alpine Malaguide Unit into two Variscan thrusted units has been demonstrated. The Silurian-Devonian succession of the Peña del Castillo belong to one Upper Variscan Unit, which backthrusts over the typical calizas alabeadas succession that in this area also constitutes one Lower Variscan Unit. The González outcrop also contitutes, very probably, a relic of a pre-Alpine unit preserved in the Epimetamorphic Alpine Malaguide Unit, but that interpretation is not still well documented. In any case, it seems that, in the studied area, the Alpine structure of the Malaguide Complex superimposes onto a pre-Alpine (Variscan) structure that overthrusted two stratigraphically different Paleozoic successions deposited in two quite different sedimentary realms within the same continental margin, in particular during the Silurian and the Devonian.

The succession recorded in the lower units (apart from the González and Peña del Castillo sections) contains turbiditic deposits as those of the Morales and Santi Petri formations, with a probably pre-Devonian and Devonian ages, respectively (although these formations have not yielded identifiable fossil remains). On the contrary, the succession recorded mainly in the Upper Malaguide Unit (that crops out to the N of the other units in present-day coordinates) includes Silurian-Devonian limestones, nodular in some places, with several types of fossil remains. Facies included in the lower Alpine Malaguide units correspond to siliciclastic-carbonatic distal (slope and basin) environments, typical of a divergent (passive) continental margin. On the contrary, pre-Carboniferous facies of the Upper Malaguide Unit point to less deep and less distal environments in the pelagic areas of the same margin during Silurian and Devonian times. Therefore, this margin opened to the SE, according to present coordinates, towards regions related to the Paleotethys ocean. Local evidences of the dismantling of more proximal (platform) environments also exist, since coral-bearing clasts have been found in conglomerates underlying the lydites of the Falcoña Formation.

In both types of Malaguide successions a maximum deepening of the basin is recorded during the Lower Carboniferous as testified by the widespread deposition of Tournaisian lydites and Visean pelagic limestones of the Falcoña Formation, which are present in all the Malaguide units and in all the sectors here studied. From this time on, the Malaguide paleomargin became tectonically

active. This fact is confirmed by the deposition of the turbiditic siliciclastic sediments of the Almogía Formation (Lower Carboniferous), traditionally interpreted as synorogenic (Variscan) Culm-type facies, and by the debris flow deposits of the Marbella Formation (Upper Carboniferous). Close to the boundary between Lower and Upper Carboniferous, short periods of tectonic quiescence probably occur, during which the deposition of the shallow-water and hemipelagic carbonates, like those dated by means of conodonts for the first time in this study took place. Carboniferous shallow-water carbonates were totally eroded and are only found as cobbles and boulders in the Marbella conglomerate Formation.

The stratigraphic succession of the Stilo Unit is thicker, more complete and less deformed in the Stilo-Bivongi area. The lower part of this sequence includes pre-Devonian pelites with intercalated metabasites, black shales and limestone beds that need further studies since they did not yielded any identifiable fossil remain. To the top it also includes limestones of Lower Devonian (dated for the first time during the course of this thesis), Middle Devonian and Upper Devonian ages, with significant lateral and vertical facies changes. Frasnian beds have been detected for the first time in Upper Devonian limestones, which allowed to verify the presence of a thin black clayed bed that coincides with the anoxic facies of the Kellwasser Event. Moreover, the lateral and vertical variability of the Devonian facies allows to characterize the pelagic areas of a continental margin with similar features to those of the more proximal environments of the paleomargin recorded in the Malaguide Complex. Lower Devonian deposits studied in the Aspromonte Massif include thicker more pelitic intercalations and carbonate facies typical of deeper enviroments. This fact points to a deposition of the southern succession of the Stilo Unit in progresivelly more distal regions of the margin. The rest of the Paleozoic sequence is characterized by Carboniferous lydites and synorogenic siliciclastic (Culm type) deposits, as in the Malaguide Complex, and points to a equivalent tectono-stratigraphic evolution.

Paleozoic successions of the Longi-Taormina Unit are mainly characterized by the presence of both calc-alkaline (acidic and intermediate) and basic alkaline vulcanism. The calc-alkaline suite is geochronologically dated as Upper Ordovician, but the alkaline basic vulcanism has not yet been dated. However, basic vulcanites that crop out in the central outcrops of the upper subunit in the Longi-Taormina Unit (Pizzo Leo area) have been interpretated as Silurian since appear below uppermost Silurian-Lower Devonian limestones well-dated by conodonts for the first time in this study. Paleozoic successions of the Longi-Taormina Unit are also characterized by lydites of probably Lower Carboniferous age and Culm type deposits.

The abundance of pre-Devonian basic volcanics and volcano-sedimentary formations and the nature of the Silurian-Devonian facies recognized within the Longi-Taormina Unit point to a palaeogeographic location of this unit in more distal realms in a continental margin than that of beds of equivalent age of the Stilo Unit. This location was closer to the intraoceanic rift axis that probably corresponded to the western sector of the Paleotethys. However, the intracontinental features of the mafic vulcanism of the Longi-Taormina Unit exclude the existence of real oceanic crust in the Paleozoic terrains here studied. This indicates that the Paleotethys was closed gradually towards its western sector.

The lateral evolution of facies in the Silurian-Devonian successions of the Malaguide Complex and of the Calabria-Peloritani Terrane reveals a proximal-distal equivalent zonation, from NW to SE andN to S (respectively and in present coordinates). Both sections constitute two transects in the same continental margin located to the N and to the NW of the Paleotethys. To sum up, Paleozoic intra-Alpine terrains studied in the Malaguide Complex and in the Stilo and Longi-Taormina units of the Calabria-Peloritani Terrane make part of the northern margin of a narrow and elongated continental crust plate (ribbon continent) incompletely detached from Gondwana in its western

end, during the Silurian-Devonian opening of the Paleotethys. This evolution fits well with that known for other nord-Gondwanic intra-Alpine regions during the opening of the Paleotethys. This caused the separation of this ribbon continent from Northern Gondwana, from which most of the Paleozoic terrains further involved in the Alpine belt are derived.

Resumen

Esta tesis estudia la estratigrafía, paleogeografía y evolución geológico-histórica del Paleozoico de los terrenos intra-Alpinos aflorantes en la Cordillera Bética y en los dominios alpinos equivalentes del Terreno Cálabro-Peloritano. El Paleozoico estudiado en la Cordillera Bética pertenece al Complejo Maláguide, el complejo tectónico superior de sus Zonas Internas. En Calabria, la unidad tectónica cuyos terrenos del Paleozoico muestran mayores semejanzas con los del Complejo Maláguide es la Unidad de Stilo, que aflora en los macizos del Aspromonte y de las Serre ocupando también la posición tectónica superior del edificio tectónico de las Zonas Internas. Ciertas afinidades maláguides, aunque también importantes diferencias, se constatan en la Unidad de Longi-Taormina, que aflora exclusivamente en los Montes Peloritanos de Sicilia, en la base y al frente del edifico tectónico de las Zonas Internas, a lo largo del contacto con dominios más externos, por tanto en una posición tectónica muy diferente de la ocupada por la Unidad de Stilo y por el Complejo Maláguide.

En todos los casos se trata de sucesiones marinas relativamente profundas, predominantemente turbidíticas terrígeno-clásticas, litológicamente monótonas, pobres en fósiles, intensamente deformadas durante los ciclos orogénicos Alpino y Varisco y, en no pocos casos, afectadas por un metamorfismo de grado bajo o muy bajo. Todo ello dificulta notablemente su caracterización litoestratigráfica, datación y correlación, así como una correcta interpretación de su significado paleogeográfico. No obstante, estas sucesiones contienen también delgadas y escasas intercalaciones carbonatadas y silíceas (radiolaríticas) que, a veces, contienen fósiles, en particular conodontos, lo que permite realizar estudios estratigráficos detallados y correlaciones fiables. La presente Tesis Doctoral se ha centrado precisamente en la caracterización litoestratigráfica, datación bioestratigráfica mediante conodontos, correlación y análisis comparado de varias sucesiones seleccionadas en las unidades tectónicas arriba mencionadas. El objetivo ha sido precisar los rasgos fundamentales de su paleogeografía y evolución geológico-histórica durante el Paleozoico, que están aún muy mal caracterizados. Para ello se han seleccionado 18 secciones en el Complejo Maláguide, 8 secciones en la Unidad de Stilo, y 6 secciones en la Unidad de Longi-Taormina para su estudio lito y bioestratigráfico. Se ha prestado especial atención a la datación de sus horizontes carbonatados mediante conodontos, que es el grupo de fósiles que se ha revelado más eficaz para cubrir los objetivos estratigráficos y paleogeográficos propuestos. Las áreas estudiadas en el Complejo Maláguide se localizan en las provincias de Almería y Murcia (Corredor de Vélez Rubio-Lorca), Granada (Zona de Cogollos Vega) y Málaga (sector de Ardales). En el S de Italia, la Unidad de Stilo se ha estudiado en dos sectores, localizados en la región de Stilo-Bivongi (Serre) y en la región de Staiti-Pietrapennata (sur del Aspromonte). La Unidad de Longi-Taormina, que aflora exclusivamente en Sicilia septentrional (Montes Peloritanos), se ha estudiado mayormente en sus afloramientos centrales y orientales.

Los estudios de campo, estratigráficos y bioestratigráficos han permitido la división del Complejo Maláguide aflorante en las tres áreas estudiadas, en tres unidades tectónicas alpinas. Estas son, de abajo hacia arriba, la Unidad Maláguide Epimetamórfica, la Unidad Maláguide Inferior y la Unidad Maláguide Superior. Estas unidades presentan, además, series paleozoicas de

características algo diferentes. En ellas se han podido caracterizar, mediante conodontos y por vez primera, horizontes del Ordovícico Superior (Hirnantiense en la región de Ardales), del Devónico (Emsiense y Famenniense) y Carbonífero Inferior (Viseense) en la Zona de Cogollos Vega, y del Carbonífero Superior (Bashkiriense) en el Corredor de Vélez Rubio-Lorca.

La sucesión litoestratigráfica de la Unidad Malaguide Epimetamórfica es claramente de tipo maláguide y semejante a la de la Unidad Maláguide Inferior. Sin embargo, la impronta metamórfica que la afecta y su posición tectónica bajo los terrenos maláguide más típicos que constituyen la Unidad Maláguide Inferior ha llevado a algunos autores a incluir parte de sus afloramientos en el Complejo Alpujárride. La Unidad Maláguide Inferior incluye la sucesión pre-carbonífera maláguide más típica, potente y extensa, constituida por las Formaciones Morales y Santi Petri, esta última caracterizada por las clásicas facies de calizas alabeadas. La serie pre-carbonífera de la Unidad Maláguide Superior es menos potente y más complicada en el detalle, tanto desde un punto de vista tectónico como estratigráfico. Está caracterizada por la presencia de horizontes estratigráficos delgados, lateralmente discontinuos e intercalados entre pelitas, formados por calizas con conodontos del Silúrico-Devónico; también por ser la única que incluye la Formación Marbella en las áreas estudiadas. Por el contrario, las formaciones del Carbonífero Inferior son semejantes en todas las unidades maláguides y están caracterizadas por las formaciones Falcoña (muy delgada, que comprende un Turnesiense formado por liditas –radiolaritas- y un Viseense constituido por calizas con conodontos) y Almogía (muy potente, turbidítica siliciclástica, en facies de tipo Culm, de edad Serpukhoviense).

Horizontes de calizas con conodontos del Silurico-Devónico también aparecen, si bien excepcionalmente, en la Unidad Maláguide Epimetamórfica (cortes del Cortijo de los González) y en la Unidad Maláguide Inferior (corte de la Peña del Castillo de Ardales). En el área de Ardales se ha podido demostrar que la Unidad Maláguide Inferior alpina se subdivide en dos unidades cabalgantes variscas, y que la sucesión siluro-devónica de la Peña del Castillo forma parte de la Unidad Varisca Superior, que cabalga sobre la típica sucesión con calizas alabeadas que, en esa zona, constituye también una Unidad Varisca Inferior. El afloramiento del Cortijo de los González es muy posible que también constituya un relicto de una unidad pre-Alpina preservado en la Unidad Maláguide Epimetamórfica Alpina, pero dicha interpretación está aún insuficientemente documentada. En todo caso, parece seguro que, en el área estudiada la estructura alpina del Complejo Maláguide se superpone a una estructuración pre-Alpina (Varisca) que puso en contacto tectónico dos sucesiones paleozoicas estratigráficamente distintas y depositadas en dos ámbitos sedimentarios bastante diferentes de un mismo margen continental, en particular durante el Silúrico y el Devónico.

La sucesión registrada en las unidades inferiores (cortes del Cortijo de los González y de la Peña del Castillo de Ardales aparte) incluye depósitos turbidíticos como los de las formaciones Morales y Santi Petri, con una edad probablemente pre-devónica y devónica respectivamente (aunque estas formaciones no han proporcionado, hasta la fecha, restos fósiles identificables susceptibles de datación). Por el contrario, la sucesión registrada sobre todo en la Unidad Maláguide Superior (que aflora al norte de las otras unidades en coordenadas actuales) incluye calizas silúricas y devónicas, en algunos lugares nodulosas, con diversos restos fósiles. Las facies incluidas en las unidades inferiores alpinas maláguides corresponden con ambientes distales siliciclástico-carbonatados de talud y cuenca, propios de un margen continental divergente (pasivo). Por el contrario, las facies pre-carboníferas de la Unidad Maláguide Superior apuntan a ambientes menos profundos y menos distales dentro de las áreas pelágicas del mismo margen durante el Silúrico y Devónico. Por tanto, dicho margen se abría hacia el SE, de acuerdo con las coordenadas actuales, hacia regiones vinculadas al océano del Paleotethys. Asimismo, existen evidencias locales del desmantelamiento de ambientes aún más proximales, correspondientes a la plataforma, ya que

se han reconocido clastos con restos de corales en conglomerados subyacentes a las liditas de la Formación Falcoña.

En ambos tipos de sucesiones maláguides se registra una máxima profundización de la cuenca durante el Carbonífero Inferior, atestiguado por el depósito generalizado de liditas del Turnesiense y calizas pelágicas del Viseense de la Formación Falcoña, presentes en todas las unidades maláguides y en todos los sectores que aquí se han estudiado. A partir de este momento el paleomargen maláguide pasó a ser tectónicamente activo. Este hecho queda confirmado por el depósito de los sedimentos turbidíticos siliciclásticos de la Formación Almogía (Carbonífero Inferior), tradicionalmente interpretados como facies sinorogénicas variscas de tipo Culm, y por los depósitos de flujos de derrubios de la Formación Marbella (Carbonífero Superior). Probablemente en torno al límite entre el Carbonífero Inferior y el Superior, ocurrieron cortos periodos de inactividad tectónica durante los cuales tuvo lugar el depósito de carbonatos de aguas someras y hemipelágicos, como los datados por primera vez mediante conodontos en este trabajo. Los carbonatos carboníferos de aguas someras fueron erosionados en su totalidad y solo se conocen como clastos en los conglomerados de la Formación Marbella.

La sucesión estratigráfica de la Unidad de Stilo es más potente, más completa y está menos deformada en el área de Stilo-Bivongi. La parte inferior de esta secuencia incluye pelitas pre-devónicas con niveles de metabasitas, pizarras negras con algunos horizontes calizos que necesitan ser estudiados más en profundidad ya que no han liberado ningún fósil identificable. Hacia arriba también incluye calizas del Devónico Inferior (datadas por primera vez durante la realización de esta Tesis), Medio y Superior con apreciables cambios laterales y verticales de facies. En las calizas del Devónico Superior se han detectado, por vez primera, niveles del Frasniense, y constatado la existencia de un delgado nivel arcilloso negro coincidente con las facies anóxicas asociadas al Evento Kellwasser. Por otra parte, la variabilidad lateral y vertical de las facies devónicas permite caracterizar las áreas pelágicas de un margen continental de características equivalentes a la de los sectores más proximales del paleomargen identificado en el Complejo Maláguide. Los materiales del Devónico Inferior de esta unidad estudiados en el Aspromonte, incluyen intercalaciones pelíticas algo más potentes y facies carbonatadas propias de ambientes algo más profundos, lo que apunta hacia un depósito de las sucesiones más meridionales de la unidad de Stilo en regiones del margen progresivamente más distales. El resto de la serie paleozoica está caracterizada por liditas carboníferas y depósitos sinorogénicos siliciclásticos (tipo Culm), al igual que en el Complejo Maláguide, y apuntan hacia una evolución tectono-estratigráfica equivalente.

Las sucesiones paleozoicas de la Unidad de Longi-Taormina se caracterizan principalmente por la presencia de vulcanismo tanto calco-alcalino (ácido e intermedio) como alcalino básico. El edificio calco-alcalino está datado geocronológicamente como Ordovícico Superior pero el vulcanismo básico alcalino no ha sido datado todavía. Sin embargo, las vulcanitas básicas que afloran en los afloramientos centrales de la subunidad superior de la Longi-Taormina (área de Pizzo Leo) han sido interpretadas como silúricas ya que aparecen por debajo de calizas del Silúrico terminal-Devónico Inferior datadas con conodontos por primera vez en este estudio. Las sucesiones paleozoicas de la Unidad de Longi-Taormina se caracterizan también por liditas de probable edad Carbonífero Inferior y depósitos tipo Culm.

La gran abundancia de formaciones volcánicas y vulcanosedimentarias básicas pre-devónicas y la naturaleza de las facies siluro-devónicas reconocidas en la Unidad de Longi Taormina apuntan hacia una localización paleogeográfica de esta unidad en dominios más distales de un margen continental que los equivalentes en edad de la Unidad de Stilo. Dicha ubicación sería cercana al eje de un rift intraoceánico que muy probablemente correspondía a los sectores occidentales del Paleotethys. Sin embargo, las características intracontinentales del vulcanismo máfico de la

Unidad de Longi-Taormina excluyen la existencia de verdadera corteza oceánica en los terrenos paleozoicos estudiados. Esto indica que el Paleotethys se cerraba gradualmente hacia sus sectores occidentales.

La evolución lateral de facies en las sucesiones siluro-devónicas del Complejo Maláguide y del Terreno Cálabro-Peloritano desvela una zonación proximal-distal equivalente, desde el NO hacia el SE y de N a S (respectivamente y en coordenadas actuales). Ambas transversales constituyen sendos transectos de un mismo margen continental localizado al N y al NO del Paleotethys. En definitiva, los terrenos paleozoicos intra-Alpinos estudiados en el Complejo Maláguide y en las Unidades de Stilo y de Longi-Taormina del Terreno Cálabro-Peloritano formaban parte del margen meridional de una placa de corteza continental estrecha y alargada ("costilla continental" o ribbon continent) incompletamente desgajada de Gondwana en su terminación occidental, durante la apertura del Paleotethys en el Silúrico-Devónico. Dicha evolución encaja bien con la que se conoce para otras muchas regiones nord-gondwánicas intra-Alpinas durante la apertura del Paleotethys. Esto determinó la separación de esa "costilla continental" desde el norte de Gondwana, de la cual derivan la mayoría de los terrenos paleozoicos que fueron más adelante tectónicamente involucrados en el cinturón alpino.

Riassunto

Questa tesi studia la stratigrafia, paleogeografia ed evoluzione geologico-storica dei terranes del Paleozoico, ripresi nel ciclo orogenetico Alpino nella Cordigliera Betica e nei domini equivalenti dell'area calabro-peloritana, che raccorda la Catena Maghrebide con quella Appenninica. I terreni paleozoici studiati nella Cordigliera Betica appartengono al Complesso Malaguide, che costituisce il complesso tettonico più elevato delle sue Zone Interne. In Calabria, l'unità tettonica, i cui terreni paleozoici mostrano maggiori somiglianze con quelli del Complesso Malaguide, è l'Unità di Stilo, che affiora nei massicci delle Serre e dell'Aspromonte e che occupa, anch'essa, la posizione più elevata dell'edificio tettonico delle Zone Interne appenninico-maghrebidi. Evidenti affinità con le successioni malaguidi, insieme a importanti differenze, sono presenti anche nell'Unità di Longi-Taormina, che affiora esclusivamente nei Monti Peloritani all'estremità nord-orientale della Sicilia, dove si ritrova alla base ed al fronte dell'edificio tettonico delle Zone Interne, a contatto con domini più esterni, e pertanto in una posizione strutturale molto diversa da quella occupata dall'Unità di Stilo e dal Complesso Malaguide.

In tutte queste unità i terreni paleozoici costituiscono successioni marine relativamente profonde, caratterizzate in prevalenza da torbiditi silicoclastiche, litologicamente monotone, povere in fossili, intensamente deformate durante i cicli orogenetici varisico e alpino e, in non pochi casi, affette da un metamorfismo di grado basso o molto basso. Tutto ciò ne rende notevolmente più difficile caratterizzazione litostratigrafica, datazione e correlazione, così come una corretta interpretazione del loro significato paleogeografico. Nonostante ciò, queste successioni contengono anche alcune sottili e scarse intercalazioni carbonatiche e silicee (radiolariti) che, a volte, contengono fossili, in particolare conodonti, che permettono studi stratigrafici dettagliati e correlazioni affidabili. Questa Tesi di Dottorato verte proprio sulla caratterizzazione litostratigrafica, la datazione biostratigrafica mediante conodonti, la correlazione e l'analisi comparata di diverse successioni scelte nelle unità tettoniche sopra ricordate. L'obiettivo è stato quello di precisare i lineamenti fondamentali della paleogeografia e dell'evoluzione geologico-storica durante il Paleozoico, che finora non sono stati mai caratterizzati. A tal scopo sono state selezionate per lo studio lito- e biostratigrafico 18 sezioni nel Complesso Malaguide, 8 sezioni nell'Unità di Stilo e 6 sezioni nell'Unità di Longi-Taormina. Particolare attenzione è stata prestata alla datazione degli orizzonti carbonatici mediante lo studio dei conosdonti, il gruppo fossile che si è rivelato il più efficace per il raggiungimento degli obiettivi stratigrafici e paleogeografici propostisi. Le aree studiate nel Complesso Malaguide sono ubicate nelle province di Almeria e Murcia (Corredor de Vélez Rubio-Lorca), Granada (Zona di Cogollos Vega) e Malaga (settore di Ardales). In Italia meridionale, l'Unità di Stilo è stata studiata in due settori, ubicati nell'area di Stilo-Bivongi (Serre) e nell'area di Staiti-Pietrapennata (Aspromonte meridionale). L'Unità di Longi-Taormina, presente esclusivamente nella Sicilia nord-orientale (Monti Peloritani) è stata studiate in prevalenza negli affioramenti centrali e orientali delle tre sub-unità in essa riconosciute.

Gli studi sul terreno, stratigrafico e biostratigrafico hanno permesso la divisione del Complesso Malaguide affiorante nelle tre aree studiate, in tre unità tettoniche alpine, denominate, dal basso in alto, Unità Malaguide Epimetamorfica, Unità Malaguide Inferiore e Unità Malaguide Superiore.

Queste unità presentano, inoltre, successioni paleozoiche con caratteri leggermente differenti. In esse è stato possibile caratterizzare, mediante conodonti e per la prima volta, orizzonti dell'Ordoviciano Superiore (Hirnantiano) nella regione di Ardales, del Devoniano (Emsiano e Famenniano) e del Carbonifero Inferiore (Viseano) nella zona di Cogollos Vega, e del Carbonifero Superiore (Bashkiriano) nel Corredor di Velez Rubio-Lorca.

La successione litoestratigrafica dell'Unità Malaguide Epimetamorfica è chiaramente di tipo malaguide e somiglia a quella dell'Unità Malaguide Inferiore. Senza dubbio, l'impronta metamorfica che la interessa e la sua posizione tettonica al di sotto dei terreni malaguidi più tipici che costituiscono l'Unità Malaguide Inferiore ha portato alcuni autori a includere parte dei suoi affioramenti nel Complesso Alpujarride.

L'Unità Malaguide Inferiore include la successione pre-carbonifera malaguide più tipica, potente ed estesa, costituita dalle Formazioni Morales e Santi Petri, quest'ultima caratterizzata dalle classiche facies di "calizas alabeadas". La successione pre-carbonifera dell'Unità Malaguide Superiore è meno potente è più complicata nel dettaglio, tanto da un punto di vista tettonico che stratigrafico. Essa è caratterizzata dalla presenza di orizzonti stratigrafici sottili, lateralmente discontinui e intercalati in peliti, formati da calcari con conodonti del Siluriano-Devoniano. Inoltre, essa è l'unica che, nell'area studiata, include la Formazione di Marbella. Al contrario, le Formazioni del Carbonidero Inferiore sono simili in tutte le unità malaguidi e sono caratterizzate dalla Formazione Falcoña, molto sottile, che comprende un Turnesiano formato da radiolariti nere (liditi) e un Viseano costituito da calcari con conodonti, e dalla Formazione di Almogia, molto potente, costituita da torbiditi silicoclastiche in facies Culm, di età serpukhoviana.

Orizzonti di calcari con conodonti del Siluriano-Devoniano appaiono anche, sia pure eccezionalmente, nell'Unità Malaguide Epimetamorfica (sezioni del Cortijo de los González) e nell'Unità Malaguide Inferiore (sezione della Peña del Castillo di Ardales). Nell'área di Ardales è stato possibile dimostrare che l'Unità Malaguide Inferiore alpina si suddivide in due unità tettoniche varisiche e che la successione siluriano-devoniana della Peña del Castillo forma una parte dell'Unità Varisica Superiore, che sovrascorre sulla típica successione con calizas alabeadas che, in questa zona, costituisce anche una Unità Varisica Inferiore. E' molto probabile che anche l'affioramento del Cortijo de los González costituisca un relitto di un'unità pre-Alpina, preservato nell'Unità Malaguide Epimetamorfica Alpina, ma questa interpretazione non è ancora sufficientemente documentata. In ogni caso, sembra sicuro che, nell'area studiata, la struttura alpina del Complesso Malaguide si sovrapponga ad una strutturazione pre-Alpina (Varisica) che ha posto in contatto tettónico due successioni paleozoiche stratigraficamente distinte e depositate in due ambienti sedimentari sufficientemente diversi di uno stesso margine continentale, in particolare durante il Siluriano ed il Devoniano.

La successione riconosciuta nelle unità inferiori (sezioni del Cortijo de los González e della Peña del Castillo di Ardales escluse) include depositi torbiditici simili a quelli delle Formazioni Morales e Santi Petri, con un'età probabilmente pre-devoniana e devonoiana, rispettivamente, sebbene queste formazioni non abbiano fornito, fino ad oggi, resti fossili identificabili e suscettibili di datazione. Al contrario, la successione presente soprattutto nell'Unità Malaguide Superiore, che affiora a nord (coordinate attuali) delle altre unità, include calcari siluriani e devoniani, a luoghi in facies nodulare, con diversi resti fossili. Le facies incluse nelle unità malaguidi alpine inferiori testimoniano ambienti distali, silicoclatici e carbonatici, di talus e bacino, propri di un margine continentale divergente (passivo). Al contrario, le facies pre-carbonifere dell'Unità Malaguide Superiore indicano ambienti meno profondi e meno distali all'interno delle aree pelagiche dello stesso margine durante il Siluriano ed il Devoniano. Pertanto, tale margine si apriva verso Sud-Est, secondo le coordinate attuali, verso regioni legate all'area oceanica della Paleotetide. Allo

stesso modo, esistono evidenze locali dello smantellamento di ambienti ancora più prossimali, corrispondenti a un'area di piattaforma carbonatica, dal momento che sono stati riconosciuti clasti con resti di coralli in conglomerati sottostanti alle liditi della Formazione Falcoña.

In entrambi i tipi di successioni malaguidi si registra il massimo dell'approfondimento del bacino durante il Carbonifero Inferiore, testimoniato dalla deposizione generalizzata di liditi del Turnesiano e calcari pelagici del Viseano della Formazione Falcoña, presenti in tutte le unità malaguidi e in tutti i settori che sono stati qui studiati. A partire da questo momento il paleomargine malaguide diviene tettonicamente attivo. Questo fatto risulta confermato dalla deposizione dei sedimenti silicoclastici torbiditici della Formazione Almogia (Carbonifero Inferiore), tradizionalmente interpretati come facies sinorogene varisiche di tipo Culm, e dai depositi da flusso della Formazione di Marbella (Carbonifero Superiore). Probabilmente intorno al limite Carbonifero Inferiore-Carbonifero Superiore, si verificarono brevi periodi di inattività tettonica durante i quali ebbe luogo il deposito di carbonati di acque basse ed emipelagici, come quelli per la prima volta datati in questo studio mediante i conodonti. I carbonati carboniferi di acque basse furono completamente erosi e la loro esistenza è provata solo dai clasti presenti nei conglomerati della Formazione di Marbella.

La successione stratigrafica dell'Unità di Stilo è più potente, più completa e meno deformata nell'area di Stilo-Bivongi. La parte inferiore di questa successione include peliti pre-devoniane con livelli di metabasiti, ardesie nere con alcuni orizzonti calcarei che necessitano di uno studio più approfondito dal momento che finora non hanno fornito nessun fossile ientificabile. Verso l'alto include anche calcari del Devoniano Inferiore (datati per la prima volta nella realizzazione di questa tesi), Medio e Superiore con significative variazioni laterali e verticali di facies. Nei calcari del Devoniano Superiore sono stati riconosciuti, anche questi per la prima volta, livelli del Frasniano, e constatato l'esistenza di un sottile livello argilloso nero, che coincide con le facies anossiche associate all'Evento Kellwasser. D'altra parte, la variabilità verticale e laterale delle facies devoniane permette di definire le aree pelagiche di un margine continentale con caratteristiche equivalenti a quelle dei settori più prossimali del paleomargine identificasto nel Complesso Malaguide. I materiali del Devoniano Inferiore di questa unità studiati nell'Aspromonte includono intercalazioni pelitiche un poco più potenti e facies carbonatiche proprie di ambienti un poco più profondi, la qual cosa indica che le successioni più meridionali dell'Unità di Stilo si sono deposte in regioni del margine via via più distali. Il resto della successione paleozoica è caratterizzata da liditi carbonifere e depositi sinorogeni silicoclastici (tipo Culm), allo stesso modo che nel Complesso Malaguide, e indicano un'equivalente evoluzione tettono-stratigrafica.

Le successioni paleozoiche dell'Unità di Longi-Taormina si caratterizzano principalmente per la presenza di vulcanismo sia calco-alcalino, acido e intermedio, sia alcalino basico. Il vulcanismo calco-alcalino è datato isotopicamente all'Ordoviciano Superiore, mentre il vulcanismo basico alcalino non è stato finora datato. Senza dubbio, le vulcaniti basiche che affiorano nell'area centrale della sub-unità superiore della Longi-Taormina (area di Pizzo Leo) sono state interpretate come siluriane dal momento che sembrano al di sotto di calcari del Siluriano terminale-Devoniano Inferiore, datati per la prima volta in questo studio per mezzo dei conodonti. Le successioni paleozoiche dell'Unità di Longi-Taormina si caratterizzano anche per la presenza di liditi di probabile età Carbonifero Inferiore e depositi di tipo Culm.

La grande abbondanza di formazioni vulcaniche e vulcano-sedimentarie basiche pre-devoniane e la natura delle facies siluriano-devoniane riconosciute nell'Unità di Longi-Taormina spingono verso un'ubicazione paleogeografica di questa unità in domini più distali di un margine continentale dell'ubicazione indicata dai depositi coevi dell'Unità di Stilo. Questa ubicazione sarebbe vicina all'asse di un rift intraoceanico che molto probabilmnente corrispondeva ai settori

occidentali della Paleotetide. Senza dubbio, i caratteri intracontinentali del vulcanismo basico dell'Unità di Longi-Taormina escludono l'esistenza di vera crosta oceanica nei terreni paleozoici studiati. Questo indica che la Paleotetide si chiudeva gradualmente verso i suoi settori occidentali.

L'evoluzione laterale di facies nelle successioni siluriano-devoniane del Complesso Malaguide e del Terrane Calabro-Peloritano svela, facendo riferimento alle coordinate attuali, una zonazione prossimale-distale simile da NW verso SE e da Nord verso Sud. Entrambe queste trasversali costituivano transetti di uno stesso margine continentale localizzato a Nord e NW della Paleotetitede. In definitiva, i terreni paleozoici ripresi nell'orogenesi Alpina, studiati nel Complesso Malaguide e nelle Unità di Stilo e di Longi-Taormina del Terrane Calabro-Peloritano, formavano parte del margine meridionale di una placca a crosta continerntale, stretta e lunga (ribbon continent), non completamente staccata dal Gondwana alla sua terminazione occidentale, durante l'apertura della Paleotetide nel Siluriano-Devoniano. Questa evoluzione si accorda bene con quel che si conosce in molte altre regioni gondwaniche intra-Alpine durante l'apertura della Paleotetide. Questo ha determinato la separazione del ribbon continent dalla parte settentrionale del Gondwana, e da esso derivano la maggioranza dei terreni paleozoici che furono in seguito inclusi tettonicamente nel sistema orogenico Alpino del Mediterraneo centro-occidentale.

CHAPTER I

General aspects

I. 1. Introduction and main objetives
I. 2. The Western Alpine Mediterranean Belt
I. 3. Materials and methods

I.1.1. Paleozoic terrains around the Western Mediterranean Sea

Paleozoic terrains around the Western Mediterranean make part of several isolated massifs that constitute the outcropping pieces of the Variscan Orogen within the Western Europe and Northwestern Africa cratons. In a lesser extent, they also outcrop as a part of the internal domains of the different mountain belts that constitute the Perimediterranean Alpine Orogen (Fig. I.1.1). The largest Paleozoic massif in Europe is the Iberian Massif, which has been practically unaffected by younger processes related to the post-Paleozoic evolution. The main geological features of the Iberian Massif, as well as of most of other European and North-African cratonic areas, were acquired at the end of the Variscan Orogeny, during the amalgamation of the Pangea supercontinent. Actually, these regions, and the epicratonic basins overlying them, mostly make part of the Alpine foreland. The present distribution of these European and North African Paleozoic terrains, however, is not that original of the Variscan Orogen immediately after the formation of Pangea, but it is related to its later disintegration. The Pangea break-up was responsible for the origin of the different european Meso-Cenozoic epicratonic basins and isolated cratonic massifs between them, as well as for the opening of the coeval Tethyan and Atlantic continental margins and oceanic basins. On the contrary, the Paleozoic terrains involved within the Western Mediterranean Alpine Orogen have not only been strongly affected by geological processes related to the Pangea formation and later break-off and opening of the Tethyan oceanic basins and margins, but also by the Alpine orogenic evolution and the later opening of the Western Mediterranean basins. Consequently, these Paleozoic terrains bear inprints of a polyorogenic evolution that renders sometimes very difficult to distinguish between pre-Variscan, Variscan and Alpine histories.

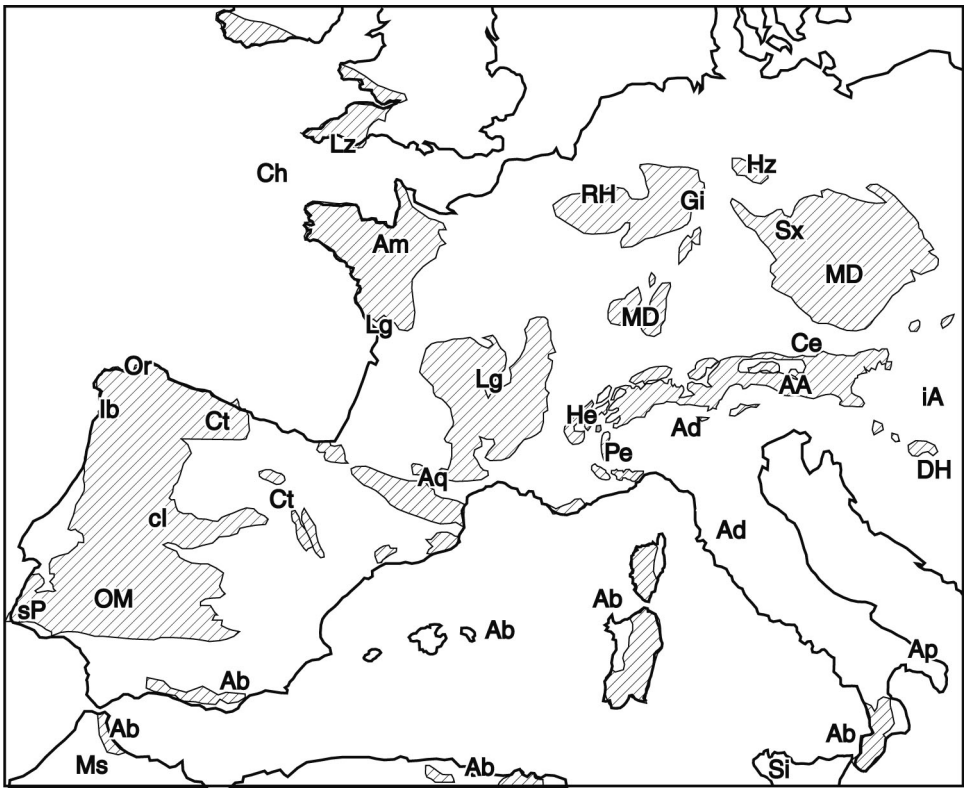


Figure I.1.1. Paleozoic terrains in Western Europe (from Stampfli et al. 2002). ***AA****, Austroalpine;* ***Ab****, Alboran (Betic-Rif-Calabria-Kabbilies-Sardinia);* ***Ad****, Adria (Tuscan Paleozoic-Southern Alps);* ***Am****, Armorica;* ***Ap****, Apulia;* ***Aq****, Aquitaine (Montagne Noire-Pyrenees);* ***Ce****, Cetic;* ***Ch****, Channel;* ***cI****, Central Iberia;* ***Ct****, Cantabria-Asturia-Ebro;* ***DH****, Dinarides-Hellenides;* ***Gi****, Giesen;* ***He****, Helvetic;* ***Hz****, Harz;* ***iA****, intra-Alpine (Tizia-Transdanubian-Bükk);* ***Ib****, allochthonous units of northwestern Iberia;* ***Lg****, Ligeria (Massif Central–South Britanny);* ***Lz****, Lizzard;* ***MD****, Moldanubian;* ***Ms****, Meseta;* ***OM****, Ossa-Morena;* ***Or****, Ordenes ophiolites;* ***Pe****, Penninic;* ***RH****, Rheno-Hercynian;* ***Si****, Sicanian basin;* ***sP****, south-Portuguese;* ***Sx****, Saxo-Thuringian.*

The progress in the knowledge about the geology of the Variscan Paleozoic massifs has been very significant in the last decades, in particular in the Iberian Massif and in the Alps. Very important contributions have been made to the stratigraphy, palaeogeography and geodynamic evolution of the pre-Alpine Paleozoic basements in some areas deformed during the Alpine cycle, such as the Iberian Chain and the Pyrenees. This is not, however, the case of the Paleozoic terrains outcropping in the different segments of the Western Mediterranean Alpine Orogen (Betic Cordillera, Moroccan Rif, Algerian Kabylias and Italian Calabria-Peloritani Arc). In these regions, most studies on Paleozoic successions deal with their pre-Alpine and Alpine tectono-metamorphic evolution, but very rarely are devoted to the study of their stratigraphy and palaeogeographic evolution. This is understandable, as most Paleozoic succession were strongly affected by tectono-metamorphic and magmatic processes related to the Variscan (Calabria-Peloritani Arc, Kabylias) and/or Alpine (Betics, Rif) orogenies, that have obliterated many -if not most- of the original sedimentary signatures of the stratigraphic successions.

This dissertation is an attempt to reduce this gap in the knowledge of the stratigraphy, palaeogeography and geological history of some of the less deformed and metamorphosed Paleozoic terrains outcropping within the Alpine chain around the Western Mediterranean in two selected areas: the Malaguide Complex of the Betic Cordillera in southern Spain, and the Stilo and Longi-Taormina units of the Calabria-Peloritani Arc, which outcrop, respectively, within the Italian Peninsula in southern Calabria and in the Sicilian Peloritani Mountains.

I.1.2. Main features of the Paleozoic palaeogeographic evolution of Northern Gondwana

It is well known that most Central and Southern European and North African Variscan regions were originally attached to the Gondwana supercontinent after its formation at the end of the Precambrian. It is also well documented that large parts of the Variscan Europe made part of several elongated continental crust pieces, actually constituting ribbon continents, that were detached from Northern Gondwana during the Paleozoic evolution. As a consequence, several oceanic basins and related continental margins were formed before their Variscan orogenic deformation. There is no consensus, however, about the number, extent and orientation of the oceanic basins and margins involved, nor about the original palaeogeographic location in Northern Gondwana of the detached landmasses during the Paleozoic northward migration of Gondwana and originally related terrains.

Different models have been proposed for the Paleozoic tectono-sedimentary reconstruction

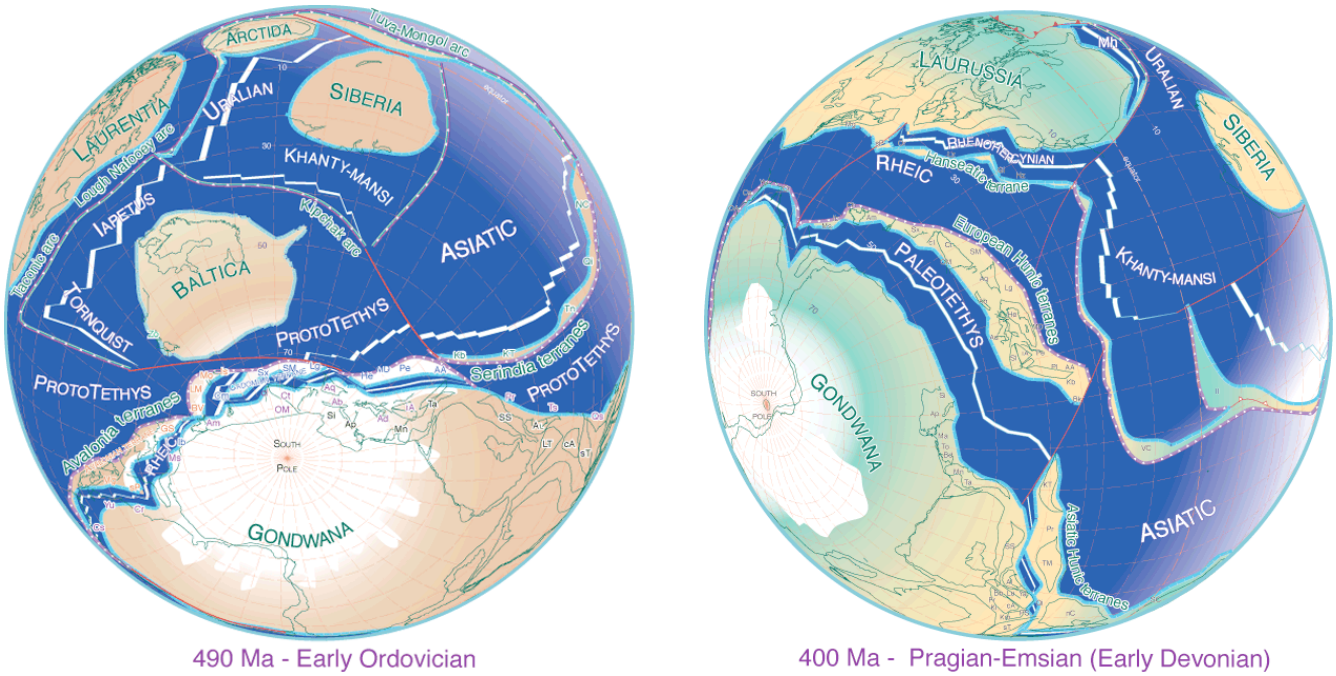


Figure I.1.2. Paleozoic palaeogeography of the Variscan european regions and opening of the Paleotethys and Rheic oceans in Western Europe according to Stampfli and Borel (2002).

and palaeogeographical evolution of the palaeomargins of Northern Gondwana, from which most European and North African Variscan regions as well as the Alpine-Mediterranean Paleozoic successions are interpreted to be derived (Robardet, 2003; von Raumer *et al.*, 2003 among many others). The development of these models is really arduous when the pre-Alpine and Alpine tectonic and metamorphic processes are superimposed, as it occurs in the Paleozoic terrains outcropping in the Western Mediterranean Alpine Belts. Therefore, such models are mostly based on studies on Variscan massifs in the Alpine forelands and in the Alps, rather than on information derived from the Paleozoic terrains of the proper Western Mediterranean Alpine Orogen.

During the last decades many authors have studied the Paleozoic palaeogeography of the European Alpine and Variscan regions, including the Iberian Massif, the Montagne Noire, the Pyrenees, the French Central Massif, the Armorican Massif, the Bohemian Massif, Sardinia, and several zones of the Alps (External Crystalline Massifs, Austroalpine Nappes of the Grauwacke Zone in the Eastern Alps, Carnic Alps, Karawanken Alps). These researches, accomplished in order to understand their pre-Variscan and Variscan evolutions, have provided us with different models (Schönlaub, 1992; von Raumer, 1998; Stampfli, 2000; Neubauer and Handler, 2000; Stampfli and Borel, 2002; Stampfli *et al.*, 2002; Gutiérrez-Alonso *et al.*, 2003; Robardet., 2003; von Raumer *et al.*, 2003, 2009; Torsvik and Cocks, 2004; Simancas *et al.*, 2005, 2009; Cocks and Torsvik, 2006; von Raumer and Stampfli, 2008; Ballèvre *et al.*, 2009; Martínez-Catalán *et al.*, 2009; Michard *et al.*, 2010, among many others). Remarkably, some of these models overbooked the role of the Intra-Alpine terrains (in the sense of Stampfli, 1996, later called Hunian and Galatian terrains by Stampfli *et al.*, 2002; von Raumer and Stampfli, 2008), to which the tectonic units studied in this dissertation are attributed, and whose importance lies in its key position in the Northern Gondwana margin during the opening and closure of the Paleotethys Ocean (Fig. I.1.2).

Other models, mainly derived from studies in the Variscan Europe, in particular in the Iberian Massif, do not recognize any solid evidence of the presence of the Paleotethys in Western Europe and North Africa, and do not consider any particular role for the Intra-Alpine terrain in their reconstructions (Simancas *et al.*, 2009, and references therein). Both groups of models differ in important aspects such as the westward continuation of the Paleotethys in Silurian to Early Carboniferous time (Fig.I.1.3). Regarding this last point, some models, mainly based on plate tectonic reconstructions (Stampfli and Borel, 2002), interpret that the Paleotethys was continuous all along the northern border of Gondwana. Nevertheless, other models, mainly

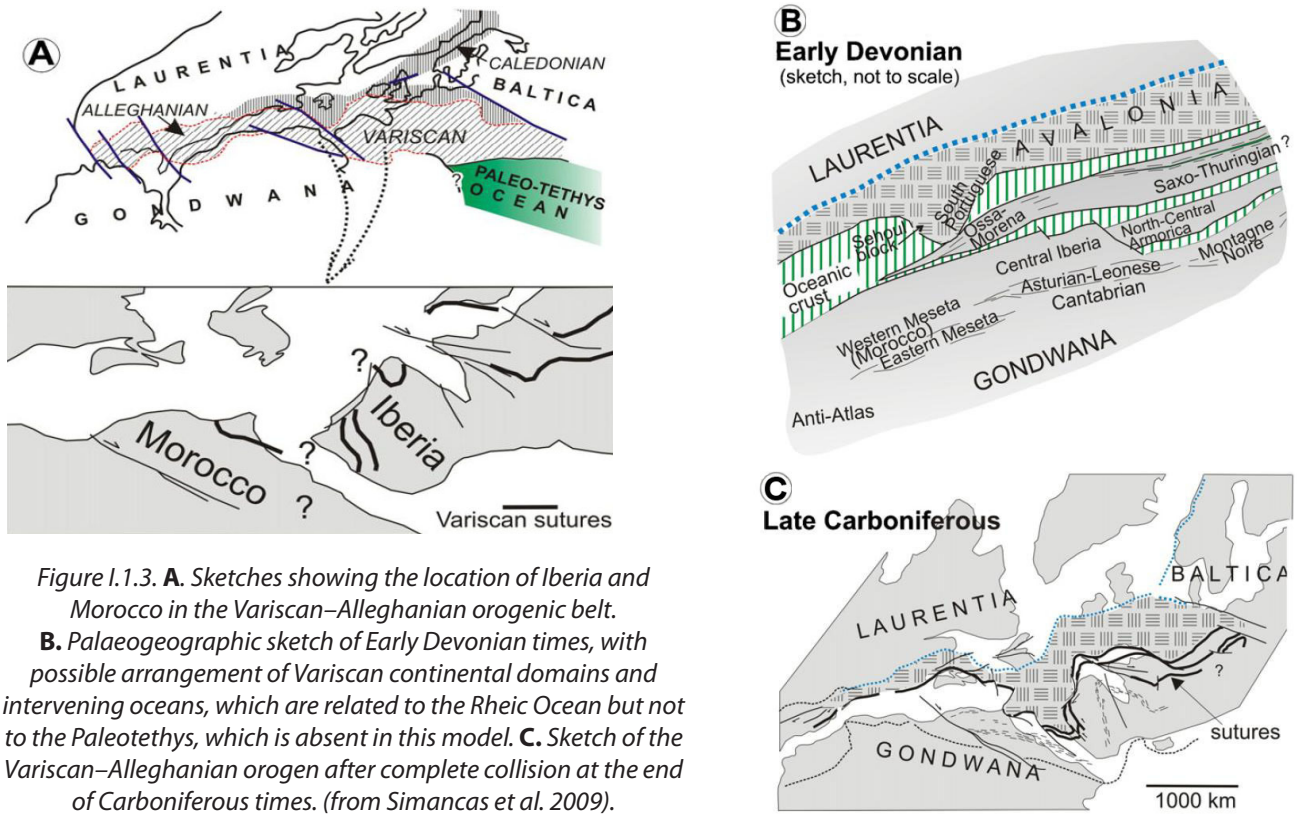


Figure I.1.3. **A.** *Sketches showing the location of Iberia and Morocco in the Variscan–Alleghanian orogenic belt.* **B.** *Palaeogeographic sketch of Early Devonian times, with possible arrangement of Variscan continental domains and intervening oceans, which are related to the Rheic Ocean but not to the Paleotethys, which is absent in this model.* **C.** *Sketch of the Variscan–Alleghanian orogen after complete collision at the end of Carboniferous times. (from Simancas et al. 2009).*

based on regional geological studies, affirm that Paleotethyan sutures cannot be found in the Variscan massifs of Iberia and Morocco, and that the ophiolite-bearing sutures present in such Variscan areas can be ascribed either to the Rheic Ocean (Simancas *et al.*, 2005, 2006, 2009; Martínez-Catalán *et al.*, 2009; Michard *et al.*, 2010) or to much older (Proterozoic) oceans (El Hadi *et al.*, 2010) but not to the Paleotethys (Fig. I.1.3). In addition, other important inconsistencies actually exist among some of these models, especially regarding to the palaeogeographic location of the Iberian Variscan terrains or the microplate character of Armorica (compare, e.g., Tait *et al.*, 2000, or Stampfli and Borel, 2002, with Robardet, 2003; Simancas *et al.*, 2009; Michard *et al.*, 2010; or Bea *et al.*, 2011).

I.1.3. Approach and aims

As mentioned before, studies on the Paleozoic stratigraphy in the Western Mediterranean Alpine belts are scarce and, in many areas, the age of only few horizons has been well constrained by biostratigraphy or geochronology. This is the case of the pre-Alpine basement rocks of the Betic Cordillera and of the Calabria-Peloritani Arc, which are affected by Variscan metamorphism and actually represent fragments of the Variscan Orogen involved in the Alpine Chain (Haccard *et al.*, 1972; Amodio Morelli *et al.*, 1976; Atzori *et al.*, 1984, 1994; Chalouan and Michard, 1990). Consequently, their correlation is rather difficult with better known Paleozoic successions of the Alps (Schönlaub, 1992, 2000; Neubauer *et al.*, 1993; von Raumer and Neubauer, 1993; Schönlaub and Heinisch, 1994; von Raumer, 1998; Schönlaub and Histon, 1999, 2000; Neubauer and Handler, 2000, among others), Minorca (Bourrouilh, 1983), Sardinia (Carmignani *et al.*, 1994; Giacomini *et al.*, 2006; Oggiano *et al.*, 2010) and, especially, the European and African Variscan Massifs (see Michard, 1976; Piqué, 1994; Dercourt, 2000; Gibbons and Moreno, 2002; Vera, 2004; Hoepffner *et al.*, 2005; or Murphy *et al.*, 2009, among many others, and references therein).

The stratigraphic features, palaeogeographic location and tectono-sedimentary evolution of both the Betic and the Calabria-Peloritani pre-Alpine basements are badly constrained and are still subject of debate. As a matter of fact, their Paleozoic stratigraphy is mainly based on integration of very limited sections carried out at different stratigraphic intervals in scattered outcrops of different Alpine tectonic units. The purpose of this study is to update and better constrain the stratigraphy and Paleozoic history of some selected outcrops of Paleozoic terrains in: 1) the Malaguide Complex of the Betic Cordillera (Vélez Rubio-Lorca Corridor in the Almería and Murcia provinces; Cogollos Vega Zone in the Granada province, and the Ardales region, in the Malaga province); 2) the Stilo Unit of southern Calabria (Serre and Aspromonte massifs); and 3) the Longi-Taormina Unit of northern Sicily (Peloritani Mountains).

The main objectives of this study are: i) to update and synthesize the Paleozoic lithostratigraphy in the different sectors; ii) to provide new biostratigraphic data on the age of the successions; iii) to correlate them, as accurately as possible, with other better known in other Alpine areas; iv) to discuss the main controls of sedimentation and the chronology of the volcanism associated with Paleozoic sedimentation; v) to discuss the fundamental features of the tectono-sedimentary evolution of the studied Paleozoic successions and the geodynamic controls on their sedimentation; and vi) to discuss their possible palaeogeographic significance and location in the frame of the Paleotethyan margins of Northern Gondwana. This attempt has been made according to recent progress in the study of other areas of the Alpine orogen where the Paleozoic evolution and palaeogeography are currently better known (e.g. Southern Alps: see Schönlaub, 1992; von Raumer, 1998; and Schönlaub and Histon, 1999, 2000; among many others).

Part of this research in both Italy and Spain has been recently published as part of my contributions to papers in collaboration with my co-workers in the scientific team of the research projects CGL2005-03887 and CGL2009-09249 (Ministerio de Ciencia e Innovación) that have supported this research. These contributions are included in this dissertation and have been completed with yet unpublished results and conclusions. The complete published contributions (articles and abstracts of congresses) are included within the appendix at the end of this memoir.

I.2. The Western Mediterranean Alpine Belt

In the Western Mediterranean, the Perimediterranean Alpine Orogen is formed by an almost closed chain, bordering the European and African coasts (Fig. I.2.1). According to their orientation and vergence, this orogen is usually divided into three segments, which are connected by two curved megastructures (oroclines or arcs). The southern segment comprises the Rif, the Kabylias and Sicily, and it connects, by means of the Calabria-Peloritani Arc, with the Apennine, which constitutes the eastern sector. The Gibraltar Arc connects the Rif and the Betic Cordillera. This latter, together with the Balearic Islands (except Minorca) constitutes most of the northwestern sector. Eastward of Majorca, the Miocene anticlockwise rotation of Sardinia and Corsica has cut off the northern segment of the orogen, which reappears with a western vergence in the Alpine Corsica, which is the southern continuation of the Alps, indeed.

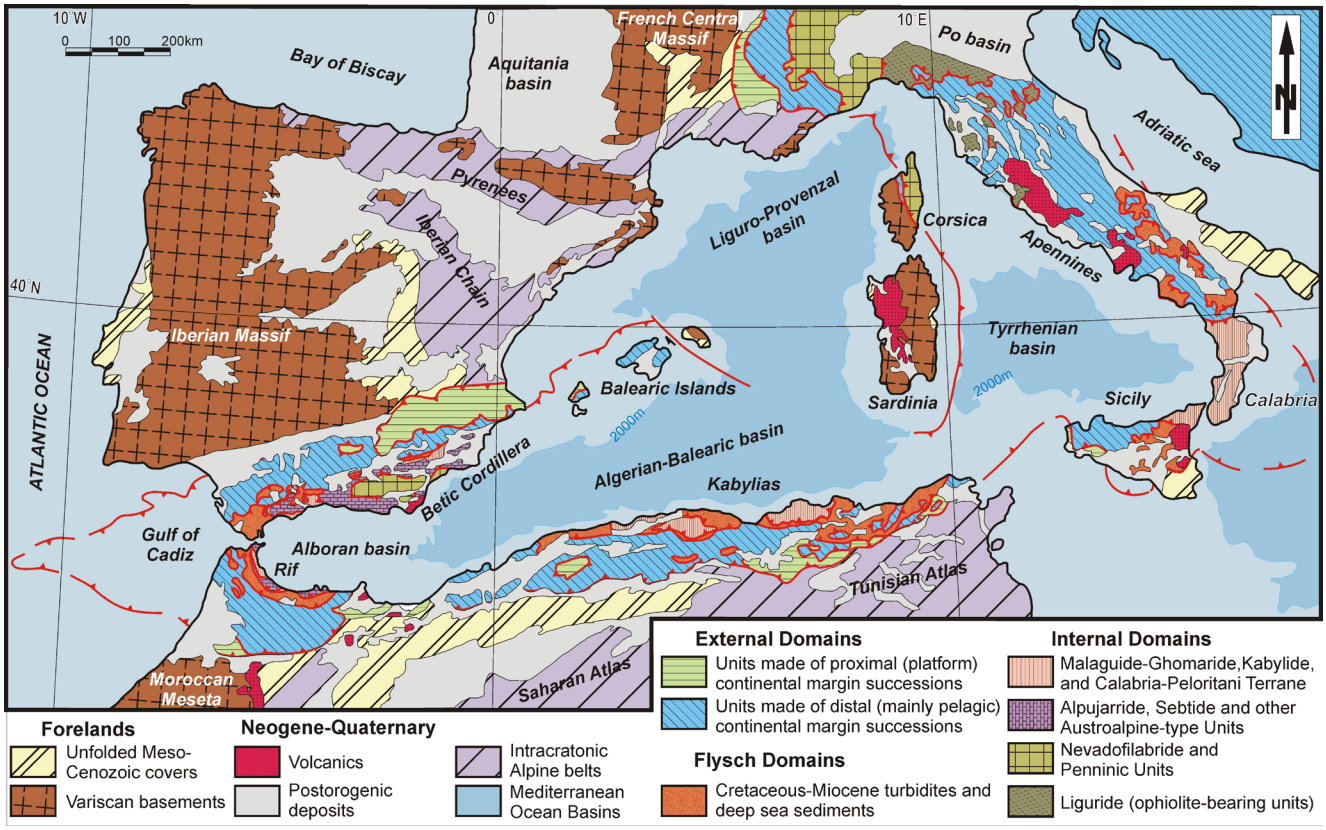


Figure I.2.1. The Western Mediterranean Orogen (from Mazzoli and Martín-Algarra, 2011, modified from Vera, 2004, and Martín-Algarra, 1987).

I.2.1. The Betic Cordillera

I.2.1.1. Major divisions, and main features of their Alpine palaeogeographical evolution

The Betic Cordillera (SE Spain) is the westernmost mountain belt of the Alpine Mediterranean orogenic system, together with the Rif in Northern Africa (Fig. I.2.1). Northward, it borders on the Iberian Massif and the Iberian Cordillera. The boundaries to the SW, S and SE coincide with the Atlantic and Mediterranean coasts. There is no doubt about the continuity of this cordillera under the sea, to connect with other Alpine domains: the North African Rif and the Balearic Islands.

As other Alpine Mediterranean chains, the Betic Cordillera is divided into Internal, Flysch, and External Domains (Durand-Delga and Fontboté, 1980; Vera and Martín-Algarra, 2004), sealed by post-Alpine extensional basins and related Neogene-Quaternary volcanism (Fig. I.2.2). All of them are made of complexes of allochthonous Alpine tectonic units, which are increasingly allochthonous from the foreland toward the hinterland, and which come from previous Meso-Cenozoic tectono-palaeogeographic domains individuated since the Permian-Triassic or Middle-Late Jurassic. In addition, some of these mega-units bear Paleozoic and older successions that also evidence a long pre-Alpine geological history of the chain.

The External and Flysch domains consist of thin-skinned tectonic units. Those of the External Domain are divided into two tectonostratigraphic subdomains named Prebetic and Subbetic (Fig. I.2.2), whose terrains constituted, respectively, the proximal and the distal parts of the Southern Iberian Meso-Cenozoic Palaeomargin (Vera, 2001). Both subdomains are made of allochthonous complexes of thin-skinned tectonic units made of Triassic to Neogene sedimentary cover successions with subordinated masses of basic volcanic and subvolcanic rocks. These successions are detached from their substratum and strongly deformed, but slightly or not affected by Alpine metamorphism (Vera, 2004).

Seismic data evidence that the basement of the External Domain consisted of thinned Southern Iberian continental crust. However, the basement of the Flysch Domain does not outcrop anywhere, as it was made of Jurassic-Cretaceous oceanic or extremely thinned continental crust, which has been subducted during the Alpine Orogeny (Durand-Delga and Fontboté, 1980). The Flysch

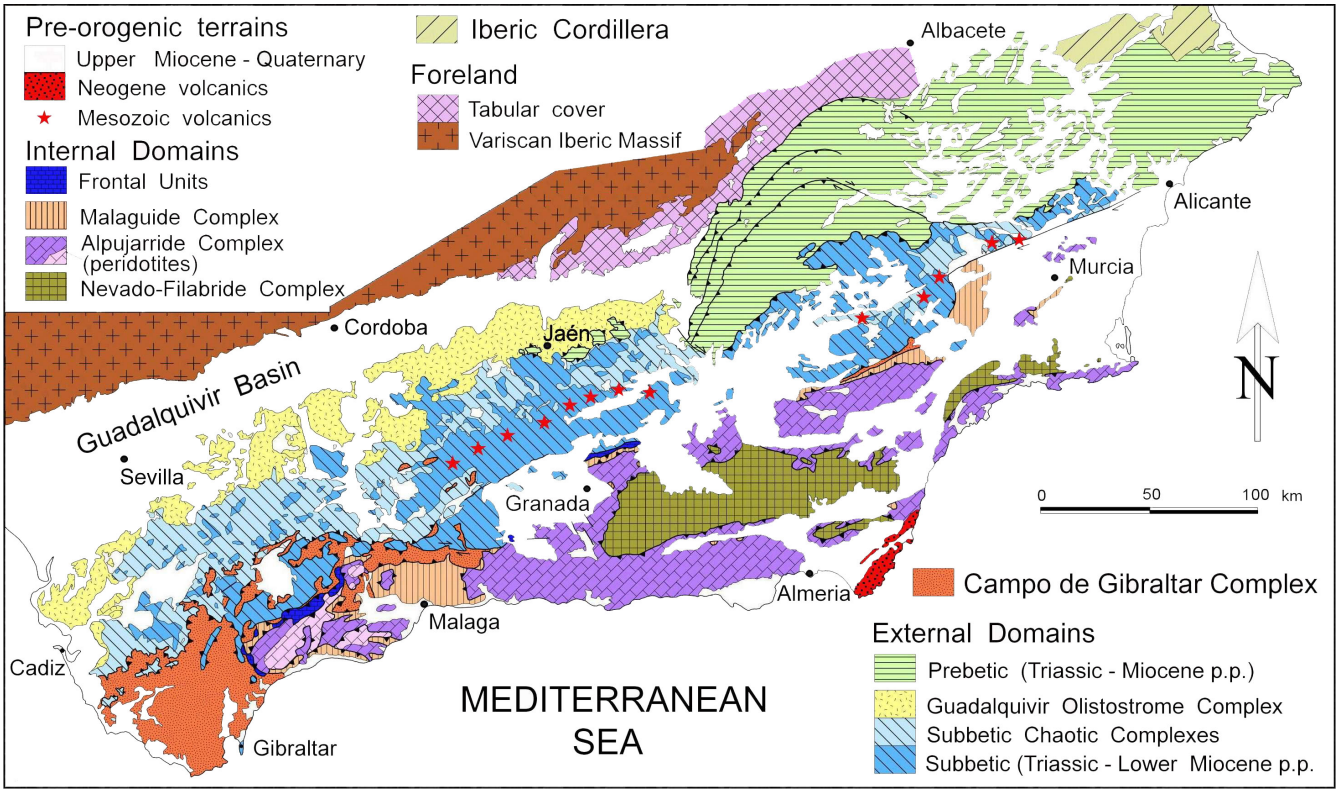


Figure I.2.2. Main divisions of the Betic Cordillera (from Perrone et al., 2006, modified from Vera, 2004, and Martín-Algarra, 1987).

Complex (Fig. I.2.2) is made of several tectonic units totally constituted by deep-marine clays and turbidites, both of carbonate and siliciclastic composition. These sediments were deposited within a deep marine basin that was opened since the Late Jurassic and that persisted up to the end of the early Miocene (Martín-Algarra *et al.*, 2004a). By the end of the Burdigalian, the cover successions of the External and Flysch domains were detached from their substratum and displaced toward the Iberian foreland, thus overthrusting the External Domain units that, at the same time, were being detached from the Southern Iberian Palaeomargin basement (Martín-Algarra and Vera, 2004b).

The Internal Domain crops out from Estepona (Málaga) to Cartagena (Murcia) and constitutes the most strongly deformed region of the Betic Orogen (Fig. I.2.2). The Internal Domain forms an antiformal stack of thick-skinned thrust nappes that bears a complex imbricated system of thin-skinned tectonic units at their front. Consequently, the Internal Domain thrust stack is subdivided, from bottom to top, into the Nevado-Filabride, Alpujarride and Malaguide complexes, plus the imbricated system of the Frontal Units (Vera and Martín-Algarra, 2004). In addition, the top of the Internal Domain is constituted by the Viñuela Group, which is made of lower Burdigalian deep marine clastic sediments unconformably deposited on top of the thrust stack (Bourgois, 1978; Aguado *et al.*, 1990; Maaté *et al.*, 1995; Martín-Algarra *et al.*, 2004c and references therein).

The Internal Domain tectonic units bear both a continental crust basement, made of Paleozoic and older rocks, and a cover made of (Permian-) Triassic to lower Miocene sediments (Vera and Martín-Algarra, 2004). In addition, subcontinental mantle peridotites are widely and exclusively present within the upper units of the Alpujarride Complex near Ronda, forming the largest peridotite outcrop in the world. Most of these units were strongly deformed and more or less affected by a Late Cretaceous to early Miocene (sometimes, and locally, also younger) Alpine metamorphism, mainly before and during (but also locally after) nappe stacking. Within the deepest part of the Internal Domain, nappe stacking probably started during the latest Cretaceous and progressed during the Paleogene (Hebeda *et al.*, 1980; Monié *et al.*, 1991; de Jong, 2003), up to the earliest Miocene (Martín-Algarra and Estévez, 1984; Martín-Martín *et al.*, 1997; Martín-Algarra *et al.*, 2000; Martín-Martín and Martín-Algarra, 2002; Vera, 2004, among many others).

In Burdigalian time, the deposition of the Viñuela Group was accounting at the same time that the thrust stack was collapsing by syn-orogenic extension and a contractional tectonic was strongly affecting the Frontal Units of the Internal Domain and the Flysch Domain units (Mazzoli and Martín-Algarra, 2011). As a consequence of the complete subduction of the oceanic substratum of the Flysch Domain and of the continental collision of the Internal Domain against the Iberian Plate, the Frontal Units were severely involved in post-Burdigalian thrusts. Moreover, this collision also determined the pluri-kilometric backthrust of the Flysch Domain units onto the Viñuela Group deposits that had just before sealed off the nappe stack of the Internal Domain units. This backthrust was followed by the antiformal folding of the Internal Domain thrust stack, and by the post-Burdigalian orogenic deformation and nappe thrusting that also affected their front as well as the External Domain tectonic units. At that time the extensional tectonics in the Internal Domain was also in progress, thus producing an extremely important thinning of the previously thickenned successions due to Alpine overthrusting (Balanyá *et al.*, 1997, among others).

From a palaeogeographical and geohistorical point of view, the Alpine evolution of the Internal Domain units of the Betic Cordillera can be summarized as follows (see also Martín-Algarra and Vera, 2004b). Together with the other Alpine Perimediterranean Internal Domains (Moroccan Internal Rif, Algerian Kabylias, and Calabria-Peloritanian Terrane) the Internal Domain units of the Betic Cordillera made part of a Triassic-Jurassic continental crust block (Mesomediterranean Block) that was rifted away from Europe and Africa during the Pangea break-off. This block was definitely detached from Iberia-Europe and Africa-Adria during the Jurassic-Cretaceous opening of narrow oceanic basins that connected the Alpine Tethys and the Central Atlantic oceans (Martín-Algarra, 1987). Since Guerrera *et al*. (1993), and according to the seminal contributions of Durand-Delga and Fontboté (1980) and Martín-Algarra (1987) this plate is known as the Mesomediterranean Microplate. In a recent paper, however, Handy *et al.* (2010) have renamed this plate as "Alcapecia", although this term can be a little bit confusing: in fact, as clearly stated by Guerrera *et al.* (1993), in late Paleogene-early Miocene time the Mesomediterranean Plate was already being deformed, and it had

been transformed into an orogenic belt that was named AlKaPeCa by Bouillin *et al.* (1986). Since Burdigalian time, and coeval to deposition of the above mentioned Viñuela Group, the AlKaPeCa (which actually had incorporated subducted and underplated pieces of the surrounding (northern) oceanic basins and margins to its nappe stack) was fragmented, and their fragments spread around the newborn Mediterranean to form the Internal Domain units of the Western Mediterranean Alpine belts.

In summary, in spite of some recent plate tectonic models that ignore the following (e.g., Schettino and Turco, 2011, among others), increasing geological evidence from the Internal Domains of the Betic Cordillera (and also from those of other Perimediterranean Alpine belts) undoubtedly demonstrates that:

1) The Mesomediterranean Block of continental crust that was detached by the Permian-Triassic continental rifting of Pangea from the main future plates of Iberia-Europe, Africa and Adria, was transformed into a independent plate during the Jurassic-Cretaceous: the Mesomediterranean Microplate.

2) During the Triassic, the central part of the Mesomediterranean Block constituted an erosion continental area that provided siliciclastic supply to syn-rift basins where continental redbeds were deposited. These redbeds are presently found within the Malaguide Complex and also, but severely modified by the Alpine metamorphism, at the base of Triassic successions of the Alpujarride Complex: they constitute the Pseudoverrucano/ Verrucano facies, which are also widespread in the Internal Domains of all the Western Mediterranean Alpine belts (Perrone *et al.*, 2006; Critelli *et al.*, 2008). These redbed successions evolved distally, and stratigraphically upwards, to shallow marine carbonate platforms with Alpine Triassic facies.

3) Since mid-Triassic up to early Miocene time the central, continental crust region of the Mesomediterranean Microplate was the site of a mainly shallow marine sedimentation, although it was also locally, and episodically, either the site of hemipelagic deposition during periods of eustatic rise or an emergent continental area during periods of sea level fall. The Malaguide Complex constituted the northern-northwestern sector of this shallow marine region.

4) Since the Early Jurassic up to the early Miocene, the above mentioned, continental to shallow marine region was totally surrounded by continental margins increasingly deeper outwards. In these margins pelagic sedimentary successions were deposited on top of shallow marine Triassic sediments. The Frontal Units of the Betic Cordillera, as well as the Alpujarride Complex and at least some units of the Nevado-Filabride Complex actually made part of the northwestern margin of the Mesomediterranean Microplate). In the Frontal Units, in particular, these pelagic successions were usually very thin and condensed, especially from the Early Jurassic up to the late Paleogene, whereas they rapidly evolved to much thicker, mainly siliciclastic turbidites in latest Oligocene and early Miocene time (Martín-Algarra, 1987), whereas in most Alpujarride and Nevado-Filabride units the original stratigraphy and facies of these successions has been strongly modified by the Alpine tectonometamorphic evolution (Martín-Algarra and Vera, 2004b).

5) The continental margins successions of the Mesomediterranean Microplate were totally bordered by deep marine basins, where sediments belonging to the Flysch Domains successions of the Perimediterranean Alpine belts were deposited. The ophiolite-bearing successions, sometimes affected by high-pressure metamorphism, which locally outcrop in some Internal Domains units of the same belts, made actually part of the oceanic crust substratum of these basins. In the Betic Cordillera relics of the basins located to the west-northwest and north of the Mesomediterranean Microplate are actually preserved, respectively, within the Campo de Gibraltar Flysch Complex and within the Nevado-Filabride Complex.

6) These Betic oceanic basins were subducted southeastwards and southwards during the early Alpine orogenic evolution, and before the continental collision of the deformed Mesomediterranean Plate, that is, the true AlKaPeCa, against the surrounding major plates and continental margins. This accounted at the end of the Burdigalian and, after the collision, the AlKaPeCa was disintegrated.

7) The western fragment of the AlKaPeCa was the Alboran Domain, which actually constitutes the Internal Domain of both the Betic Cordillera and the Rif. The rapid westward migration of the Alboran Domain, and its persistent collision against Iberia and North Africa during post-Burdigalian time,

determined the bending of its western part to form the Gibraltar Arc (which is a middle-late Miocene but not older tectono-palaeogeographic feature as considered by most models), and the detachment and orogenic deformation of the Meso-Cenozoic successions deposited in the South-Iberian Palaeomargin to constitute the External Domain of the Betic Cordillera.

In brief, to discuss the pre-Alpine evolution of the pre-Mesozoic rocks that constitute wide parts of the Internal Domain of the Betic Cordillera and of other areas of the Perimediterranean Alpine Orogen, which is the main objective of the present study, one must take into account that the present distribution of the Paleozoic facies belts outcropping around the Western Mediterranean results not only from the pre-Alpine palaeogeographic and orogenic evolution but also from their younger Alpine pre-orogenic and orogenic evolution. Actually, the correlation of the studied Paleozoic successions, and their interpretation within a coherent pre-Alpine palaeogeographic and palaeotectonic scenario, cannot be correctly performed if they are not restituted, accordingly to the Alpine pre-orogenic and orogenic evolution, to their original location at the end of the Paleozoic before the early Mesozoic Pangea disintegration.

I.2.1.2. Paleozoic successions of the Betic Internal Domain

Pre-Mesozoic rocks are absent in the Frontal Units of the Internal Domain but widely present in the Malaguide, Alpujarride and Nevado-Filabride complexes. Nevertheless, in the two later complexes the Paleozoic and older successions are exclusively made of metamorphic rocks (mainly derived from sediments but also from both plutonic and volcanic rocks). Consequently, very rare evidence of the presence of fossils has been noticed in these complexes, and studies on the stratigraphy of these successions are very rare, and mainly based on structural and petrological analyses. Nonetheless, Devonian fossils and scarce remains of Precambrian acritarchs have been discovered, respectively, within rocks of the lowermost units of the Nevado-Filabride Complex in the Cartagena (eastern Betic Cordillera: Lafuste and Pavillon, 1976) and Sierra de Baza areas (Gómez-Pugnaire *et al.*, 1982).

In the Alpujarride Complex the existence of pre-Mesozoic fossils has never been mentioned, although Triassic and younger fossils have been found and a reasonably good knowledge of the Mesozoic stratigraphy is available. Nonetheless increasingly important information about the Paleozoic and older history of their successions is emerging from the study and dating of zircons extracted from polymetamorphic rocks, mainly from the uppermost Alpujarride units. On the contrary, Paleozoic (meta-)sedimentary rocks are

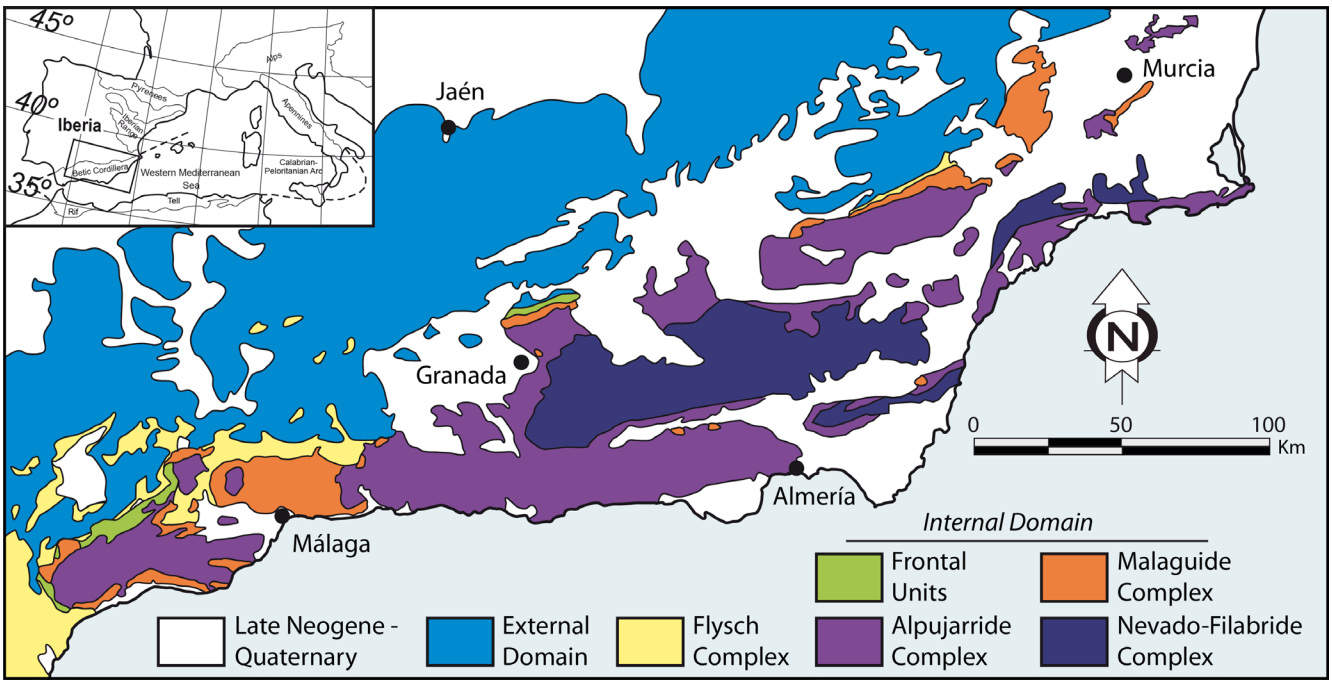


Figure I.2.3. The Internal Domain of the Betic Cordillera (modified from Vera, 2004).

widespread in the Malaguide Complex, and, in spite of strong Alpine (and pre-Alpine) tectonic deformation, its stratigraphic frame is much more better known than that of the other two Internal complexes.

The Malaguide Complex (Fig. I.2.3) makes the topmost part of the Internal Domain, and it is slightly affected by the Alpine metamorphism. It thrusts widely over the Alpujarride Complex although, in most places, the original contact has been modified by younger extensional tectonics. These terrains were firstly defined by Blumenthal (1927) as Betic of Malaga. Later, Durand-Delga (1968) introduced the term Malaguide, when he recognized its presence to the south of Sierra Nevada (see also Bodenhausen *et al.*, 1967). The Malaguide Complex mainly outcrops in the western Costa del Sol, the Montes de Málaga and the Ardales area, but also to the north of Granada (Cogollos Vega Zone), the Vélez Rubio-Lorca Corridor, Sierra Espuña and other sierras of Murcia (Blumenthal, 1927, 1928, 1949; Blumenthal and Fallot, 1936; Fallot, 1945, 1948; Soediono, 1970; Geel, 1973; Herbig, 1983, 1985; Martín-Algarra, 1987, Martín-Martín, 1996, among many others).

The Malaguide Complex is made of a Paleozoic basement formed, mainly, by siliciclastic sediments of deep marine facies and affected by the Variscan and Alpine orogenies, but its low metamorphic grade makes possible to recognize the pre-Alpine history of this domain. The Triassic to lower Miocene cover lies, unconformably, on this Paleozoic substratum, it is usually rather thin (except in Sierra Espuña), and it is mainly formed by continental (Triassic) to shallow marine or hemipelagic sediments with many stratigraphic gaps (Jurassic-Cretaceous) and mainly warm water (tropical) carbonate facies. Uppermost Oligocene and lower Miocene syn-orogenic sediments usually form thick siliciclastic successions eroded from, mainly, sedimentary successions with lithofacies equivalent to those of the Malaguide Paleozoic substratum, and/or from equivalent pre-Mesozoic sedimentary and metamorphic/magmatic rocks presently outcropping in other units of the Western Mediterranean Alpine belts (Martín-Algarra *et al.*, 2000). This youngest part of the Malaguide succession is Upper Oligocene-Aquitanian in age (Ciudad Granada Group: Soediono, 1971; Martín-Algarra, 1987; Martín-Martín, 1996) and shows a wide variety of facies (from continental to turbiditic). Its stratigraphic and petrographical analysis allows to know the palaeogeographical evolution of the Betic Internal Domain in a time period (25-20 Ma) when the Alpujarride and Nevado-Filabride complexes were located in deeper structural levels and, consequently, were being affected by a strong Alpine deformation and metamorphism. At the same time, the Malaguide Complex or the Frontal Units were located in the superstructure of the orogenic belt being the site of the thick siliciclastic sedimentation within moderately deep marine basins that constitutes the Ciudad Granada Group (Martín-Algarra *et al.*, 2000). In addition, the stratigraphic study of the Malaguide Paleozoic successions allows a much better understanding of the pre-Alpine history of the pre-Triassic successions of the Alpujarride and Nevado-Filabride complexes.

I.2.1.3. Previous stratigraphic studies on the Malaguide Paleozoic

The Paleozoic rocks of the Malaguide Complex (Fig. 1.2.4) are mainly slates and greywackes, predominantly of turbiditic origin and poor in fossils. They are crosscut in some places by early Miocene subvolcanic dykes (Torres-Roldán *et al.*, 1986). They also include conglomeratic intercalations, thin siliceous beds (originally radiolaritic cherts) and pelagic calcareous levels bearing fossils, mainly conodonts (Kockel, 1960, 1963; Kockel and Stoppel, 1962; Rodríguez-Cañero, 1993, O'Dogherty *et al.*, 2000, among others).

In the Vélez Rubio area, these Paleozoic rocks were included by Soediono (1971) and Geel (1973) in the Piar Formation. Later on, Martín-Algarra (1987) changed the stratigraphic rank of this lithostratigraphic unit into a group and, also using data from previous authors (in particular from Michelau, 1942; Mon, 1969, 1970; and Herbig, 1983, 1984), he subdivided the Piar Group in five formations named, from bottom to top: Morales, Santi Petri, Falcoña, Almogía and Marbella. These deposits range from a pre-Silurian age to Late Carboniferous. During the course of this dissertation, a Hirnantian (Late Ordovician) horizon has also been found (Rodríguez-Cañero *et al.*, 2010, see also appendix I). This is the oldest horizon well dated by means of fossils in the Malaguide Complex. Many Silurian to Carboniferous levels have been dated by means of radiolarians and conodonts in the pelagic intercalations (Herbig, 1984, 1985, 1986, Rodríguez-Cañero, 1993, 1995, O'Dogherty *et al.*,

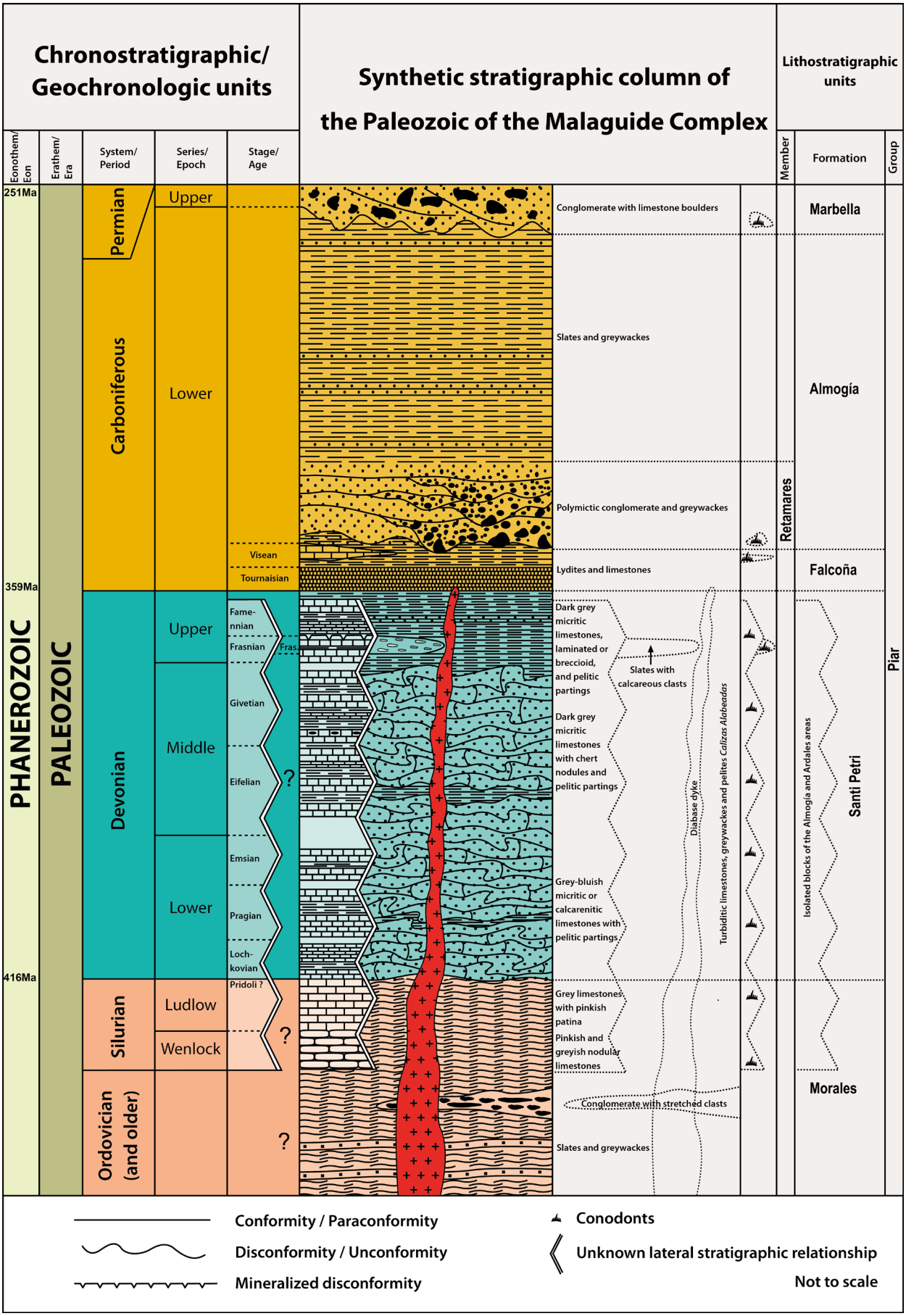


Figure I.2.4. Synthetic Paleozoic stratigraphy of the Malaguide Complex (modified from Martín-Algarra et al., 2004b).

2000). However the correlation between these well-dated levels and the classical formations in the Piar Group is only possible in a rough way because they appear either forming isolated blocks limited by tectonic contacts or as olistoliths included within deep marine clastic deposits.

The Morales Formation (Mon, 1969, 1971) is the oldest formation of the Malaguide Complex despite it still remains undated. This formation consists of a very thick pelite-psammite succession with a progressive downwards increase of the metamorphism, so that its deepest part can be very difficult to distinguish from Alpujarride pre-Mesozoic rocks (Benamocarra Unit in the eastern *Montes de Malaga* area: Macaione, 2011, and references therein). The Morales Fm. is made up of fine-grained slates and shales alternating with compact quartz-feldspathic and micaceous sandstone beds (including, occasionally, some microconglomerate levels) where foliation sometimes has not completely removed the original features so that, locally, some sedimentary structures (cross lamination and graded bedding) can be observed. In the deepest levels of the Morales Fm. a dark dotted structure is widespread (Gálvez and Orozco, 1980). It is formed by graphite aggregates of sericite, sometimes with relics of metamorphic minerals (andalusite and garnet: Tubía, 1985). Below these levels, Aldaya *et al.* (1979) and Elorza (1982) defined the Benamocarra Unit, which they attribute to the Alpujarride Complex. Recent studies (Ruiz-Cruz, 1997; Ruiz-Cruz and Nieto, 2002), however, suggest a lithologic and metamorphic continuity with the base of the Malaguide Complex, although an extensional detachment surface seems always to be present between them (Macaione, 2011). A metric to decametric, laterally discontinuous horizon made of strongly deformed clast-supported conglomerates, and unformally known as *Conglomerado de Cantos Estirados* (that is, conglomerate with stretched clasts) appears quite frequently towards the top of the Morales Fm.; it is composed predominantly by stretched and flattened siliciclastic and locally carbonatic clasts of pebble and cobble size. This formation, and locally higher formations, is crossed by millimetric to decametric basic subvolcanic dykes, characterised as arc tholeiites, and dated as early Miocene (Torres-Roldán *et al.*, 1986; see also Platt *et al.*, 2003).

The Santi Petri Formation (Michelau, 1942, Mon, 1971) has a very variable thickness because of tectonics but it usually reaches several hundreds of metres. Its facies ranges from fine-grained limestones to calcareous greywackes, usually strongly folded and known as *Calizas alabeadas* (Orueta, 1917). This turbiditic formation commonly presents Bouma sequences with a truncated base, proving a distal turbiditic environment of deposition. In spite of lacking fossils and because of its position below well-dated levels (Falcoña Fm.), a pre-Carboniferous age (probably Devonian) is commonly accepted.

In the topmost part of the Santi Petri Fm. a few-metres thick pelitic-arenitic interval, marks a gradual lithological transition to the Falcoña Formation (Herbig, 1983). This transitional interval was firstly identified in the western Malaguide outcrops by Kockel and Stoppel (1962), who named them as *Serie Rana*. In the Falcoña Fm., two members can be distinguished. The lower member, radiolaritic in origin, is made up of very thin-bedded dark cherts usually known as lydites. They probably had an original thickness of a few tens of metres in most cases but they are usually strongly thinned due to intense disharmonic folding, flattening and shearing, and they are also frequently transformed into brownish and orangish –less commonly, whitish- quarzites (Orozco and Gálvez, 1979; Gálvez and Orozco, 1979, 1980). Cherts in equivalent stratigraphic position form a well defined lithostratigraphic horizon in Lower Carboniferous successions of many European regions. As in many of these areas, in the Malaguide Complex this horizon has been dated as Tournaisian by means of radiolarians (O'Dogherty *et al.*, 2000). The upper member of the Falcoña Fm. is calcareous, and it consists of micritic limestones (sometimes also including chert nodules), which have been dated by means of conodonts as upper Visean (Rodríguez-Cañero and Guerra-Merchán, 1996; Navas-Parejo *et al.*, 2006, 2008, see appendix I).

In many areas, a several-metres-thick shaly interval with some fine-grained and thin-bedded sandstone, marks a gradual transition from the upper beds of the Falcoña Fm. to the overlying Almogía Formation. Nonetheless, in other areas, this formation seems to be unconformable on older beds, which are usually much more strongly deformed. This makes frequently the Falcoña Fm. to disappear under the lower and frequently coarser-grained beds (Retamares member) of the Almogía Fm. The latter consists of a turbiditic megasequence of greywacke and pelite, with some conglomerate intercalations usually thinning- and fining-upwards. It is commonly several hundreds

of metres thick, and shows typical Variscan Culm facies. The lowest part of this formation is the Retamares member (Kockel and Stoppel, 1962; Herbig 1984, Rodríguez-Cañero, 1993), which is made up of medium- to coarse-grained and thick to very thick bedded greywackes and conglomerates that were deposited in proximal, frequently channelled, deep marine fan environments. The conglomerates contain limestone clasts that have provided Middle to Upper Devonian conodonts and other fossils (Buchroitner *et al.*, 1980; Herbig, 1984). The rest of the formation comprises brownish to olive-green shales and thin-bedded and fine-grained greywackes, sometimes bearing undeterminable plant remains, typical of turbiditic distal environments.

The top of the Almogía Formation is usually an unconformity surface below the Permian?-Triassic redbeds with Pseudoverrucano lithofacies included in the Saladilla Formation (Roep, 1972). Nonetheless, the youngest Paleozoic deposits of the Malaguide Complex belong to the Marbella Formation (*Conglomerado de Marbella*). The latter rests disconformably on the Almogía Fm., from which it can be easily distinguished by its massive appearance and much coarser grain size. In addition, the Marbella conglomerates usually contain big clasts (cobbles and boulders up to several metres large) of granite and gneiss, and especially of shallow-water limestones bearing Carboniferous fossils (Herbig, 1984, 1986; Herbig and Mamet, 1985). The deposition of the Almogía and Marbella formations demonstrate the synorogenic erosion of Paleozoic successions that were being deformed during the Variscan Orogeny. This erosion should be related to emergence of pre-Carboniferous metamorphic and granitic terrains that, in some areas were covered by Carboniferous shallow marine carbonate platforms.

For this dissertation field observations have been made in three of the most important Malaguide outcrops: the Ardales area (Málaga province), the Cogollos Vega Zone (Granada province) and the Vélez Rubio-Lorca Corridor (Almería and Murcia provinces). A total of sixteen sections in these areas have been measured and sampled for conodont studies, as they include limestone intervals. The lithostratigraphic succession in many other outcrops without biostratigraphic information as they lack pelagic limestone horizons, were also checked from its integration within the stratigraphic scheme of the Malaguide Paleozoic.

I.2.2. The Calabria-Peloritani Terrane

I.2.2.1. Main features and major divisions

The Calabria-Peloritani Terrane (CPT) is a fault bounded, exotic terrane located at the intersection between the NW-SE trending Southern Apennines to the north, and the E-W trending Sicilian Maghrebides to the south (figs. I.2.1 and I.2.5). The CPT borders at north on the fold-and-thrust belt of the Apennines and at south on the Sicilian Maghrebides, and constitutes an arc-shaped orogenic belt, frequently known as Calabria-Peloritani Arc. This morphological concept is, however, rather confusing when refers to their geology and, consequently, Bonardi *et al.* (2001) substituted it by the concept terrane. According to the latter authors, the north end of the CPT structurally overrides the uppermost nappes of the Southern Apennines, whereas its south border is thrust over the uppermost nappes of the Sicilian Maghrebides.

In spite of being part, geographically, of the Apennines-Maghrebides system, the CPT is geologically quite different from the surrounding chains. Firstly, the CPT is characterised by thick-skinned Alpine continental crust nappes, showing a pre-Triassic crystalline basement, which thrust over HP ophiolite-bearing thrust sheets, the latter being pieces of the Western Tethys; the Apennines and the Maghrebides, on the contrary, are made exclusively of thin skinned thrust sheets. Moreover, the CPT shows evidence of pre-Neogene tectonics, whereas the Apenninic and Maghrebian thrust sheets were emplaced during the Neogene. All these nappes have been strongly thinned by syn- to late-orogenic extensional tectonics.

The CPT is bound by the Sangineto (to the north) and the Taormina (to the south) tectonic lines. The nature and features of these boundaries are still subject of debate because of the lack of good seismic profiles imaging them at depth. Based on field observations, the Sangineto line seems to be a complex fault system with a left-lateral

strike-slip component and the Taormina line a NNE-dipping thrust fault, later affected by dextral strike-slip motion (Somma, 2006). To the west, the CPT continental crust attenuates gradually, and laterally changes to the Pliocene-Quaternary oceanic crust of the Tyrrhenian seafloor. To the east, the CPT is bound by the Ionian subduction zone, where the relict Mesozoic Ionian oceanic crust is subducted northwestward underneath the CPT. Actually, a large portion of the CPT is covered by the Ionian Sea.

Lugéon and Argand (1906) and Limanowsky (1913) first pointed out the existence of nappes within the CPT. Until then, the basement of the CPT was interpreted as the culmination of an autochthonous Apenninic-Sicilian chain (Cortese, 1895; De Lorenzo, 1896). Taking into account the Argand's conception of the Alps, and on the base of extensive field and analytical work, Quitzow (1935) interpreted southern Calabria as a Hercynian massif with the structure of a large mushroom fold. During the Alpine orogeny, this massif behaved as the root zone for nappes verging both to the north (northern Calabria) and the south (Peloritani Mountains). Since Ogniben (1960), however, the CPT has been interpreted as a stack of rootless nappes.

Compared to other Alpine Perimediterranean orogenic sectors, the Southern Apennines (southern Italy) and the Sicilian Maghrebids (central-western Sicily) constitute the External Domain of the Western Mediterranean Alpine belt in Sicily and in the Italian peninsula (Fig I.2.5; Bonardi *et al.*, 2001, 2005). The

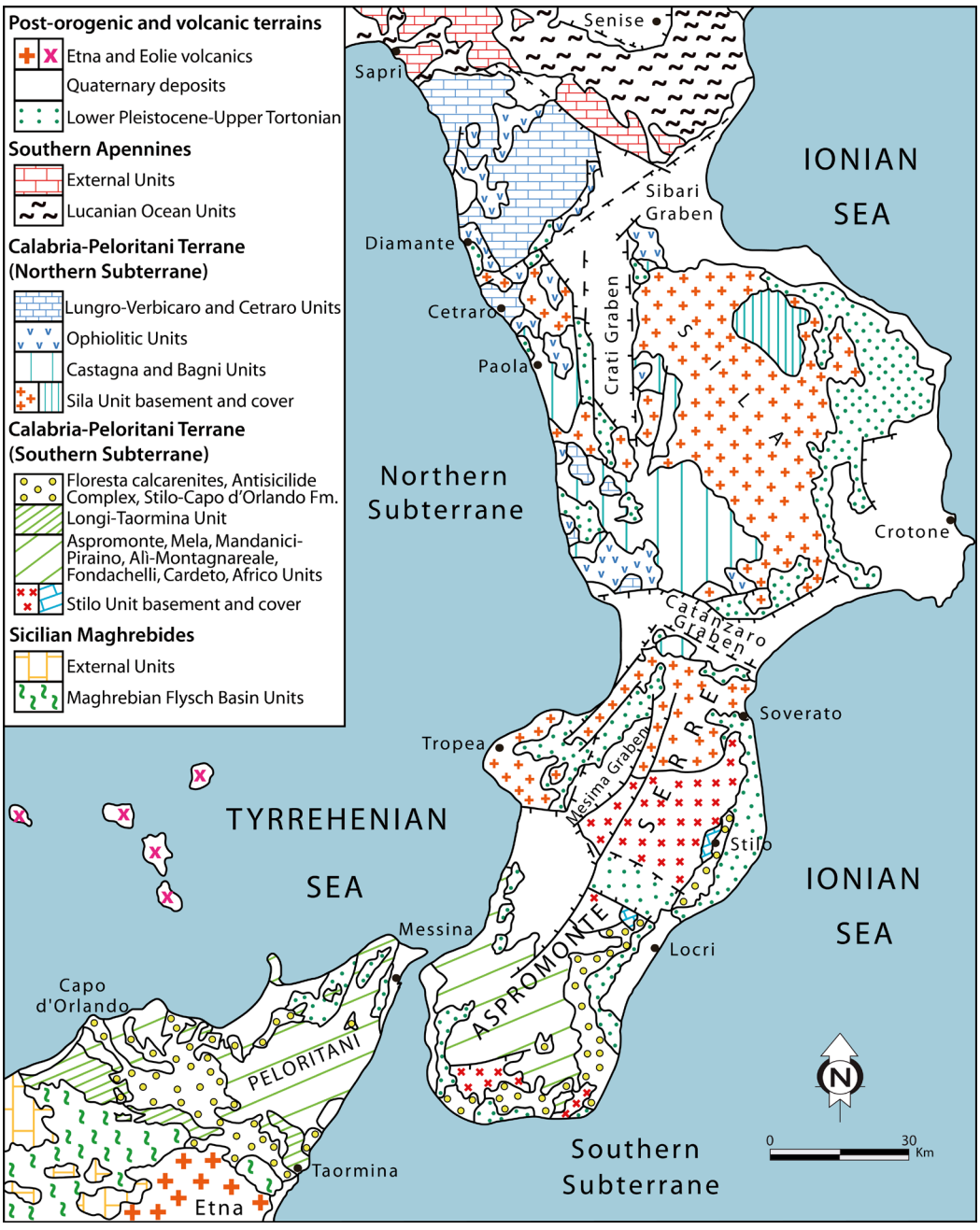


Figure I.2.5. The Northern and Southern subterranes of the Calabria-Peloritani Terrane (modified from Perrone et al, 2006).

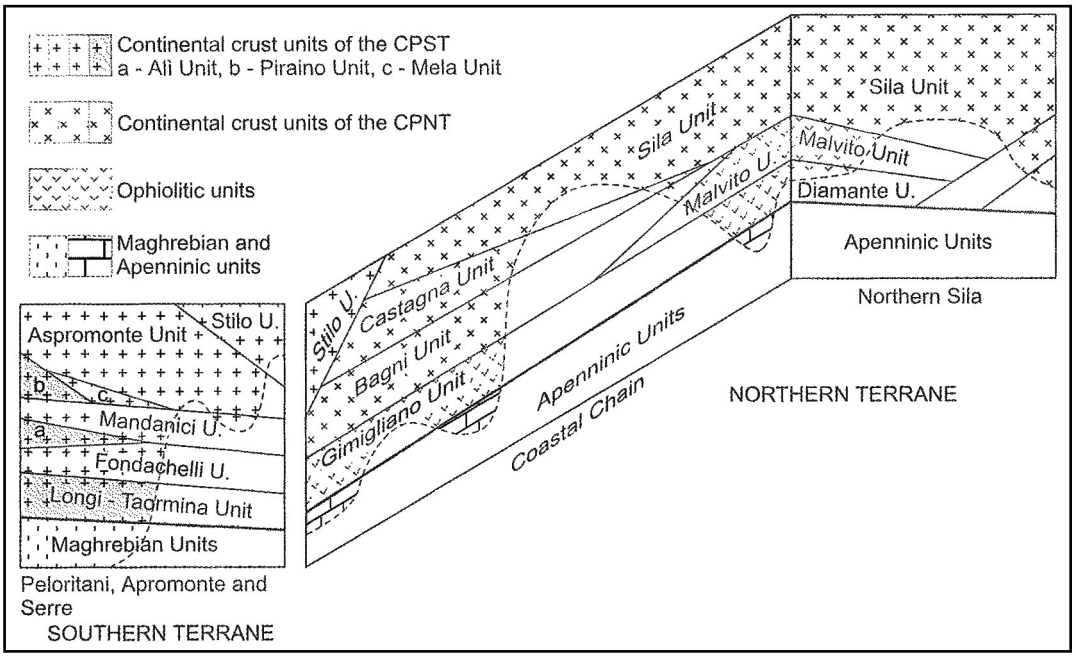


Figure I.2.6. Geometrical relationships between the CPT tectonic units. Dashed line: boundary between the outcropping and buried (white) portion of the chain (from Bonardi et al., 2004).

CPT, however, plays the role of the Internal Domain, as it is characterized by a thick-skinned tectonics involving pre-Mesozoic crystalline basement that shows evidence of pre-Neogene deformation, and by Alpine metamorphism that, in some units, locally affects the pre-Mesozoic continental basements rocks and, in other units also, the Meso-Cenozoic cover. These features are absent in the Southern Apennines and in the Sicilian Maghrebides, which, as already said, are made only of thin-skinned thrust sheets that were tectonically emplaced exclusively during the Neogene.

Because of the difficulty of correlating the tectono-stratigraphic evolution of the units of the CPT in northern Calabria with those of southern Calabria and Sicily, the CPT has been subdivided into two sectors (figs. I.2.5 to I.2.9) with different nappe architecture (Fig. I.2.6) and Late Cretaceous-Miocene tectonostratigraphic and metamorphic evolution (Bonardi *et al.*, 1980, 2001, 2005 and references therein). The Northern Subterrane ranges approximately from the Sangineto line to the Mesima graben, represents a fragment of an Europe-vergent eo-alpine (latest Cretaceous-early Paleogene) chain, and includes the Sila and the Catena Costiera massifs (Fig. I.2.5). The Southern Subterrane includes the Serre and Aspromonte massifs (Fig. I.2.7) and the Peloritani Mountains (Fig. I.2.8).

The tectonic stack in the Northern Subterrane is formed, from bottom to top, by the Diamante-Terranova, Malvito, Gimigliano and Monte Reventino ophiolitic units, and by the, Bagni, Castagna and Sila continental crust units (Fig. I.2.6). Apart from the Sila Unit, which is constituted by a thick pre-Mesozoic succession essentially made of Variscan metamorphic and magmatic rocks covered by a locally thick succession of unmetamorphosed Mesozoic and Cenozoic rocks, both the basement (either formed by pre-Mesozoic continental, or by Mesozoic oceanic crust) and the Meso-Cenozoic cover of the tectonic units underlying the Sila Unit, are affected by Alpine metamorphism (Fig. 1.2.9).

The Southern Subterrane ranges from the Soverato-Mesima valley alignment (to the N) in Calabria to the Taormina Line (to the S) in Sicily (figs. I.2.7 and I.2.8). In the Aspromonte-Serre massifs (Fig. I.2.7), the Alpine tectonic stack is formed by four units, that are, from base to top: Africo, Cardeto, Aspromonte and Stilo units (Bonardi *et al.*, 2001; 2002; figs. I.2.6 and I.2.9). In the Peloritani Mountains (Fig. I.2.8), the Alpine unit stack is composed, from base to top, by the Longi-Taormina, Fondachelli, Alì-Montagnareale, Mandanici, Piraino, Mela and Aspromonte units (Bonardi *et al.*, 1976, 2001; Messina *et al.*, 2004; figs. I.2.6 and I.2.9). This subterrane is characterised by an Africa-verging Alpine nappe pile, which is unconformably capped by the Burdigalian Stilo-Capo d'Orlando Formation (figs. I.2.7 and I.2.8), and it is formed only by continental crust nappes and slices, which override the Maghrebian Monte Soro Unit, which makes part of the Flysch Domain in Sicily together with other units, some of them bearing Numidian Flysch deposits (Nicosìa and Monte Salici Units).

Contrary to the Northern Subterrane, in the Southern Subterrane of the CPT there is no evidence for a Late Cretaceous to early Paleogene tectono-metamorphic evolution. There, the nappe stack was established, and underwent a very rapid exhumation and erosion, in Aquitanian–Early Burdigalian time. This Miocene orogenic evolution is testified by the Aquitanian terrains involved in the nappe stack, and by the Burdigalian Stilo-Capo d'Orlando Formation sealing the nappe structure, which is exclusively present in this sector (Fig. I.2.5). The deposition of this formation predates the collision of the Calabria-Peloritani Southern Subterrane against the outermost Sicilian Magrebide belt, as well as the back-thrusting of ocean-derived sediments of the Maghrebian Flysch Domain, that constitute the so-called Antisicilide Complex. The latter is finally covered, unconformably, by the Langhian-Serravallian Calcarenite di Floresta Formation, and by younger upper Miocene, Pliocene and Pleistocene, mainly marine deposits.

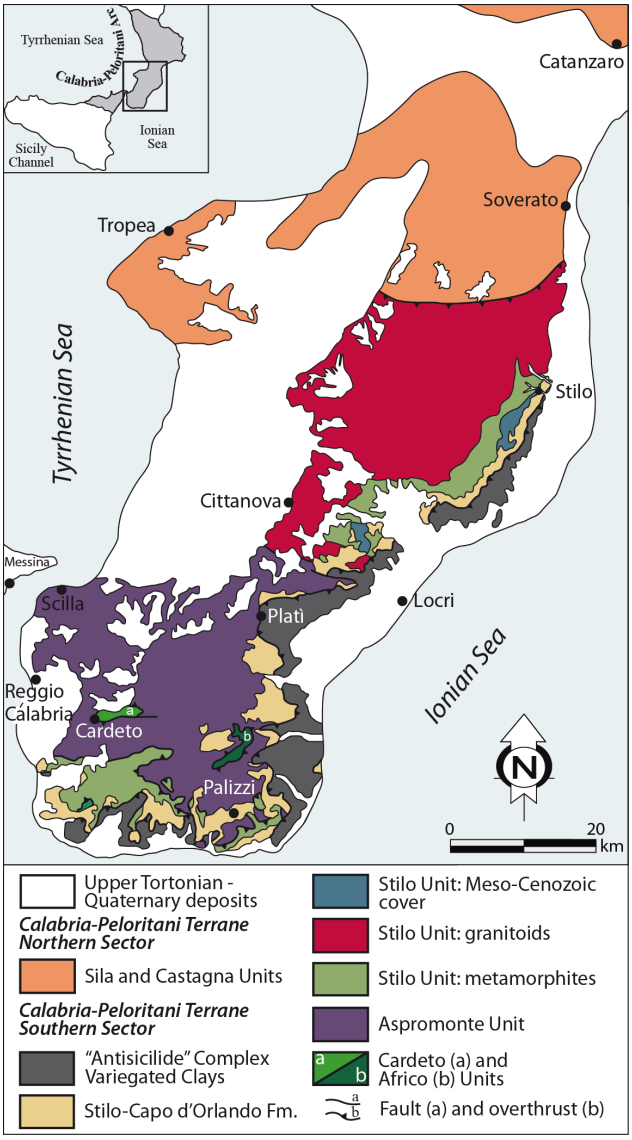


Figure I.2.7. Tectonic sketch of the Southern Subterrane of the CPT (Aspromonte-Serre massifs) (modified from Navas-Parejo et al., 2009a).

The tectonic units of the Southern Subterrane are mainly made up of pre-Triassic rocks affected by Paleozoic magmatic (both effusive and intrusive) and metamorphic events. Unmetamorphosed Mesozoic-Cenozoic sedimentary covers overlie crystalline rocks, except in the Aspromonte and the Mela units, which are devoid of sedimentary rocks (Fig. I.2.9). In addition, the Aspromonte, Mandanici and Ali units are locally affected by Alpine metamorphism mostly in greenschits facies but locally reaching the amphibolite facies (Aspromonte Unit).

As concerns the age of the pre-Alpine metamorphism, it is Late Carboniferous (314 Ma Rb/Sr radiometric age, Bonardi *et al.*, 2008) and related to the Variscan orogeny. A Visean metamorphism (323-347 Ma K/Ar radiometric age, Guerrera *et al.*, 1999) characterises the Longi-Taormina Unit (Fig. I.2.9). A late-orogenic felsic plutonism (300-290 Ma Rb/Sr radiometric age, Rottura *et al.*, 1993; Ayuso *et al.*, 1994) is widespread in the Stilo and Aspromonte units (Fig. I.2.9). Particularly, in the crystalline rocks of the different units of the Peloritani Mountains, Variscan metamorphism gradually evolves, from anchimetamorphism-greenschist facies in the lowermost unit (Longi-Taormina Unit; Somma *et al.*, 2005b) up to the amphibolitic facies in the uppermost units (Mela and Aspromonte units; Messina *et al.*, 1997; De Gregorio *et al.*, 2003).

The age of the Alpine metamorphic overprint is Late Oligocene in age (22-28 Ma Rb/Sr radiometric age, Bonardi *et al.*, 1984, 2002, 2008). This event ranges from greenschist to amphibolite facies, and locally affects the Aspromonte Unit, both in the Peloritani Mts. and the Aspromonte-Serre massifs. An Alpine metamorphic overprint and metamorphism in subgreenschist facies (26 Ma Rb/Sr radiometric age, Atzori *et al.*, 1994) has also been recognized in the basement and the Mesozoic rocks of the Mandanici Unit. Analogously an Alpine anchimetamorphism has been found in the Alì Unit (Somma *et al.*, 2005a, Somma and Martín-Rojas, 2011, and references therein). The Alpine metamorphism is, however, absent in the Stilo, Mela and Longi-Taormina units (Somma *et al.* 2005b).

As concerns the stratigraphy of the pre-Mesozoic basements of the CPT, it is badly known and summarized, together with that of younger deposits and of other neighbour units in Fig. I.2.9.

In the Northern Subterrane fossiliferous beds have been reported only in the Sila Unit in the studies of Bouillin *et al.* (1984, 1987). In these papers acritarchs should indicate a Cambro-Ordovician age. Later, fossiliferous levels never have been recognized in the Sila Unit, where Paleozoic magmatic and strongly metamorphic rocks predominate. In the Southern Subterrane, thick and well preserved Paleozoic successions have been recognized exclusively in the Stilo Unit of the Aspromonte-Serre massifs and in the Longi-Taormina Unit of the Peloritani Mountains. However, fossiliferous beds are very scarce, and consequently, the knowledge of the stratigraphic features of the Paleozoic successions is scanty and poorly defined.

In the 1970s, studies on the pre-Alpine basements of the CPT showed the only presence of Middle-Upper Devonian dacryoconarids and conodonts in metacalcareous lenses of the Stilo and Longi-Taormina units (Afchain, 1970; de Capoa-Bonardi, 1970; Lardeaux and Truillet, 1971). Later, successions starting in the Cambro-Ordovician and ending with Carboniferous Culm facies were recognised in the Stilo and Longi-Taormina units (Majesté-Menjoulas *et al.*, 1984, 1986; Bouillin *et al.*, 1987, 1989). However, only Cambro-Ordovician and Middle-Upper Devonian layers have been dated by means of a very few specimens of acritarchs and conodonts, respectively. Consequently, the reconstructed successions and their ages are questionable because they are based mainly on facies similarities and/or field characters of volcanics occurring in them. Actually, acceptable reconstructions of the stratigraphic successions lack and, consequently, different tectono-sedimentary reconstructions have been proposed for these fragments of the Variscan orogeny involved in the Alpine chains (Fig. I.2.10). In fact, the location of the CPT basements in the Variscan chains of the central-western Mediterranean area is subject of debate (Spalletta and Vai, 1989; Acquafredda *et al.*, 1989, 1994).

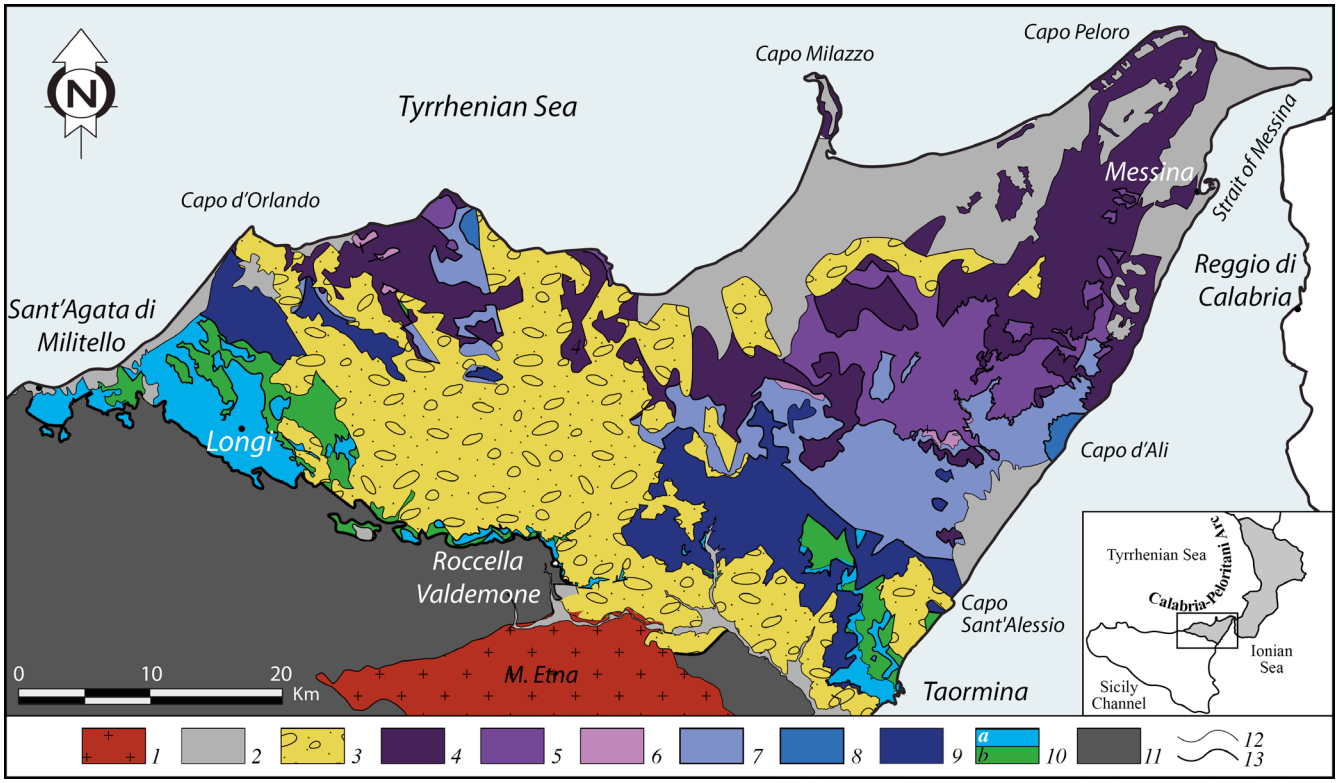


Figure I.2.8. Tectonic sketch of the Peloritani Mountains (modified from Somma, 2006). ***1****, Etna volcanics (Pleistocene-Holocene);* ***2****, Holocene and Pleistocene-Miocene deposits;* ***3****, Calcareniti di Floresta (Serravallian-Langhian), "Antisicilide Variegated Shales" (Upper Cretaceous-Paleogene) and Stilo-Capo d'Orlando Fm. (Burdigalian);* ***4****, Aspromonte Unit: Variscan medium-high grade metamorphics with Pre-Variscan granulitic relics, intruded by Late-Variscan plutonics. Localised Alpine medium- to low-grade metamorphic overprint;* ***5****, Mela Unit: Variscan medium-grade metamorphics with Eo-Variscan eclogitic relics;* ***6****, Piraino Unit: Jurassic-Triassic sedimentary cover and Variscan low- to medium-grade basement;* ***7****, Mandanici Unit: Mesozoic sedimentary cover and Variscan low- to medium-grade basement;* ***8****, Alì-Montagnareale Unit: Mesozoic sedimentary cover and Variscan low-grade basement;* ***9****, Fondachelli Unit: Mesozoic sedimentary cover and Variscan low-grade basement;* ***10****, Longi-Taormina Unit:* ***a****) Meso-Cenozoic sedimentary cover;* ***b****) Variscan very low- to low-grade basement;* ***11****, Sicilian Maghrebid units (Middle Miocene-Upper Jurassic flysch deposits);* ***12****, Stratigraphic contact;* ***13****, Tectonic contact.*

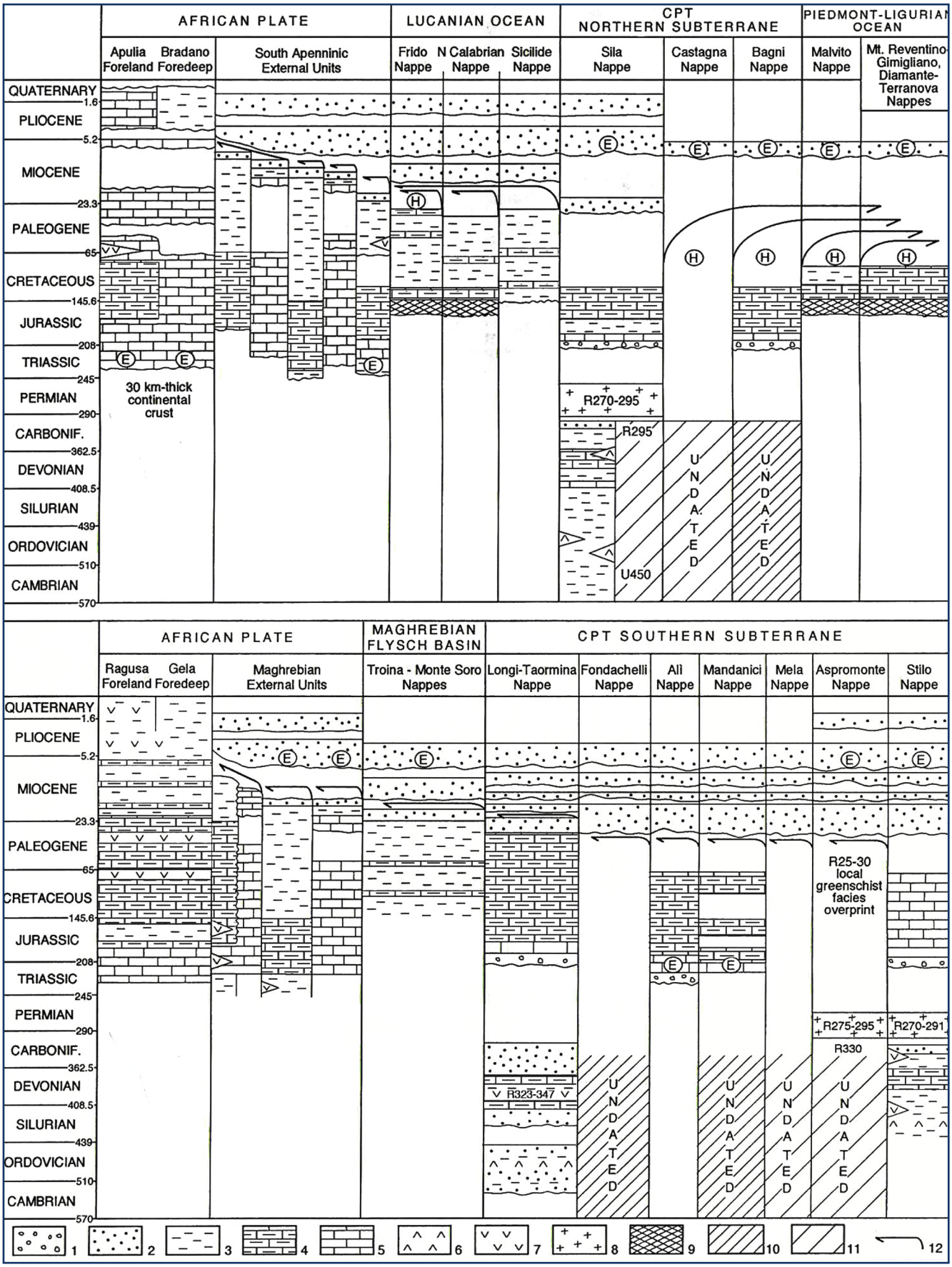


*Figure I.2.9. Synthetic stratigraphy of the Calabria-Peloritani Terrane (slightly modified from Bonardi et al., 2001). **1**, Continental clastics; **2**, Marine siliciclastics (mainly turbidites, both siliciclastic and carbonatic); **3**, Pelagic marls and clays, locally including carbonate turbidites; **4**, Pelagic carbonates and radiolarites; **5**, Shallow marine carbonates; **6**, Felsic volcanics; **7**, Mafic volcanics; **8**, Plutonic rocks; **9**, Ophiolitic sequences; **10**, Low-grade metamorphic rocks; **11**, Medium and high-grade metamorphic rocks; **12**, Accretionary thrusts; **E**, Evaporites; **H**, HP/LT Alpine metamorphism. Radiometric ages: R (Rb/Sr) and U (U-Pb).*

I.2.2.2. General features of the Stilo and Longi Taormina units

The Stilo Unit is the highest thrust-sheet in the nappe stack of the Calabria-Peloritani Southern Subterrane, and crops out only in southern Calabria (Fig. I.2.7). It consists of both a pre-Alpine basement and a Meso-Cenozoic cover, this latter severely reduced by erosion (Fig. I.2.9). The basement is made up of Variscan metamorphites in which a gradual transition from amphibolite facies to greenschist facies is observed (Bonardi *et al.*, 1984). The low-grade metamorphites consist mainly of phyllites, metarenites and slates, with minor metabasites, and recrystallized limestones and black cherts. The metamorphic succession is intruded by late-Variscan calc-alkaline plutonic rocks. The sedimentary cover starts with continental redbeds a few metres thick, followed by Upper Triassic?–Lower Jurassic? dolostones and limestones. The succession continues up to the Jurassic–Cretaceous boundary with platform facies, consisting of limestones and calcareous breccias, with frequent lacunae marked by bauxitic clay layers. Upper Cretaceous–Eocene marine rocks have never been documented. The earliest Oligocene is represented by marsh marls, marly limestones and algal limestones, followed by a layer of red clays and by Upper Rupelian–Chattian calcarenites and calcareous breccias with Lepidocyclina, rich in metamorphic and granitoid clasts fed by the underlying basement. The succession ends with a few metres of Aquitanian turbiditic sandstones. Alpine metamorphism is lacking, and the whole succession is unconformably covered by the Stilo-Capo d'Orlando Fm.

Most of the studies on the Paleozoic low-grade metamorphic succession of the Stilo Unit have been made in the Stilo-Bivongi area, where it constitutes the so-called Bivongi series. Six main lithostratigraphic units have been distinguished there (Bouillin *et al.*, 1987, 1989). The lowest layers are represented by several hundred metres of low-grade silvery phyllites and grey metarenites with intercalated lenses of metabasites, that Bouillin *et al.* (1989) attributed to the Cambro-Ordovician. Its top is marked by a thin and laterally discontinuous horizon of black metapelites with some associated metalimestone lenses, with less than 10 m of total thickness and both considered to be of Silurian age. These rocks are followed by grey metapelites, alternating with thin- and fine-grained metarenite beds, more than 200 m thick, locally including greenish quartzites and, towards its top, thin-bedded limestones with some dacryoconarids, indicating a Devonian age (Afchain, 1970). Upwards, several strongly folded and laterally discontinuous limestone lenses, a few metres to several tens of metres thick and showing nodular structure and reddish colours (*griotte* facies), have provided late Famennian conodonts (de Capoa-Bonardi, 1970; Bouillin *et al.*, 1987) and constitute the only well-dated horizon in this succession before the present study. These limestones are topped by few metres of thin-bedded cherts, considered to be Early Carboniferous in age. Finally, the succession ends with more than 100 m of dark slates alternating with turbiditic sandstones, sometimes bearing plant remains and showing the classical Carboniferous Culm facies.

The Longi-Taormina Unit (LTU) crops out along a WNW-ESE trending belt, stretched from Sant'Agata di Militello in the Tyrrhenian coast to Taormina in the Ionian coast (Fig. I.2.8). The LTU successions, mainly in the central area, are covered by the late-orogenic deposits of the Stilo-Capo d'Orlando Fm and, therefore, their outcrops are divided in a western sector (from Acquedolci to Banco di Formisia-Pagliara), a central sector (from Pizzo del Moro to Malvagna) and an eastern sector (from Novara di Sicilia to Taormina). The LTU overthrusts onto the Monte Soro Unit of the Maghrebian Flysch Domain along the frontal thrust of the Peloritani Mountains tectonic building, the Taormina Line (Bonardi and Giunta, 1982; Giunta, 1991). The LTU, on its turn, is overthrust by the Fondachelli Unit (Fig. I.2.9).

The LTU involves both Paleozoic sedimentary and volcanic protolithes mildly affected by Variscan metamorphism, and a thick Mesozoic-Cenozoic sedimentary cover (figs. I.2.9 and I.2.10). The basement is maily formed by Ordovician acidic to intermediate volcanites, followed by undated (probably Silurian-Devonian) basic volcanics (Fig. I.2.11). These volcanic rocks are intercalated within low-grade dark coloured metapelites and metapsammites where Cambro-Ordovician acritarchs have been locally found, including Devonian limestone lenses and probably Carboniferous cherts and pelites towards the top (Fig. I.2.11). Radiometric data on the pre-Triassic rocks testify for a Late Ordovician acidic volcanism (456-452 Ma U/Pb age, Trombetta *et al.*, 2004) and a Visean metamorphism (323-347 Ma K/Ar age, Guerrera *et al.*, 1999), that reached the subgreenschist to greenschist facies (Cirrincione *et al.*, 1999).

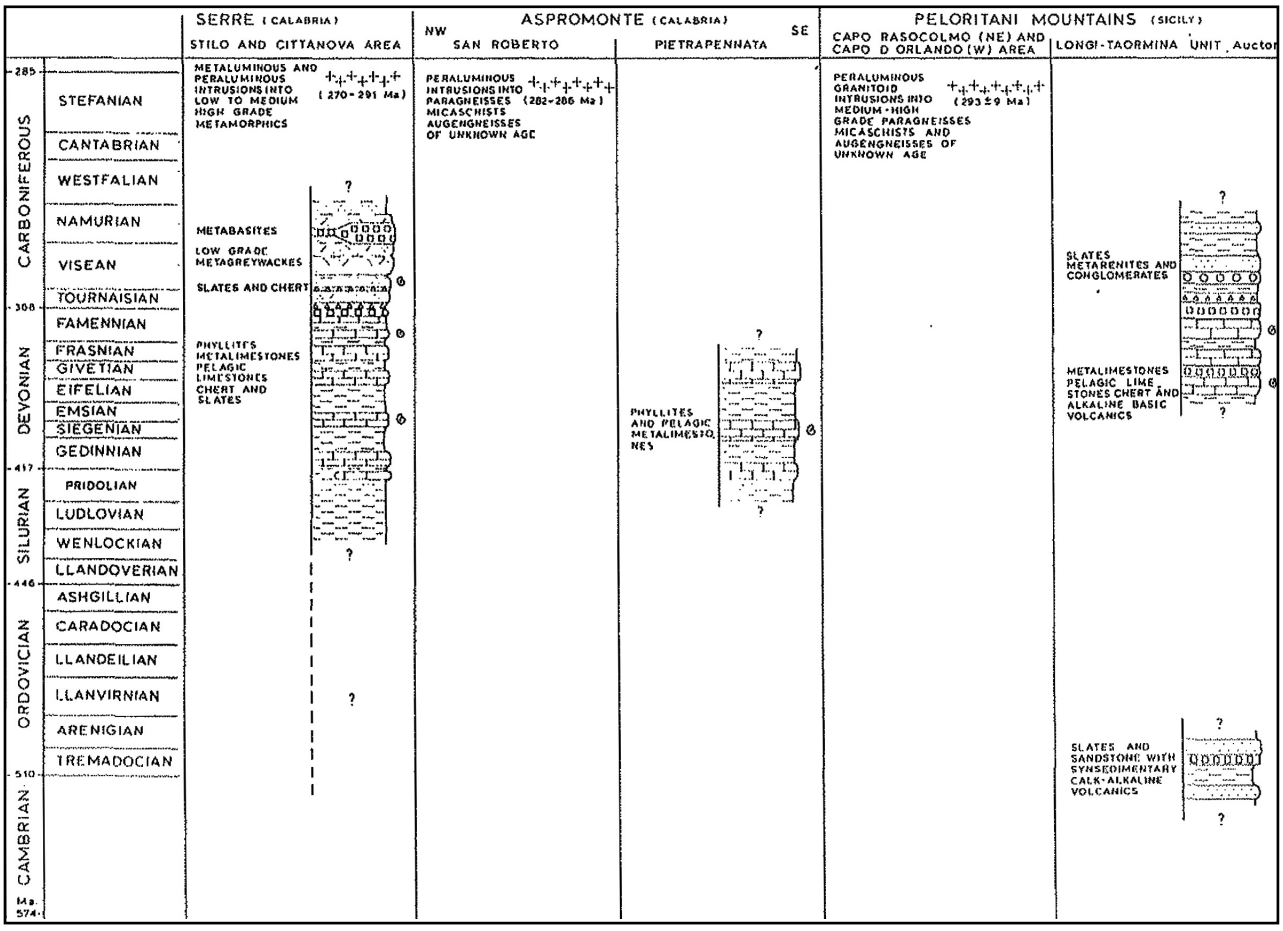


Figure I.2.10. Stratigraphic correlation of the Stilo and the Longi-Taormina units (from Spalletta and Vai, 1989).

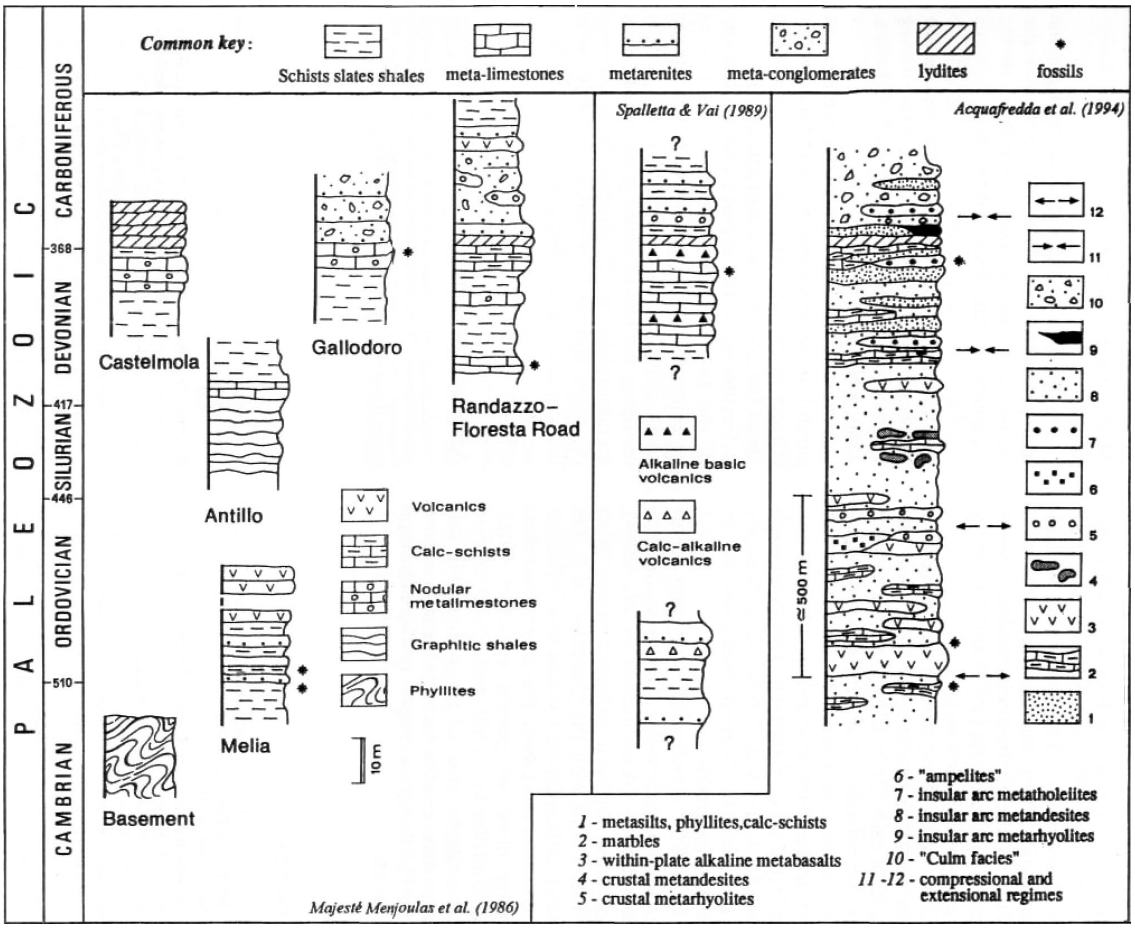


Figure I.2.11. Different stratigraphic columns of the Paleozoic basement of LTU (Guerrera et al., 1999).

Somma (1999) (Taormina-Roccella area)	Ogniben (1960)	Duee (1961)	Truillet (1969-70)	Lentini and Vezzani (1975)	Amodio-Morelli *et al.* (1976)
Longi-Taormina Unit first order unit (Paleozoic basement and Mesozoic-Cenozoic cover) Constituted by:		Chaine Calcaire	Chaine Bordiere		Longobucco-Longi-Taormina Unit
Second-order unit **Rocca Licopeti-Pizzo Leo** (basement and cover with evolution from continental to neritic platform (lowermost Lower Jurassic); from platform to slope (uppermost Lower Jurassic); and from slope to shallow marine (Ypresian, Eocene)	Layer of Longi	Galati Unit	Taormina Unit	Taormina Unit	tectonic slice
Second-order unit **Taormina** (basement and cover with evolution from continental to neritic platform (lowermost Lower Jurassic); from neritic platform to basin (middle Lower Jurassic); and from basin to abyssal plane (Tithonian-Eocene)					
Second-order unit **Capo Sant'Andrea-Gallodoro** (basement and cover with evolution from continental to neritic platform (lowermost Lower Jurassic); from neritic platform to sea-mount (middle Lower Jurassic); and from sea-mount to bathyal plain (Tithonian-Eocene)		Longi Unit	Longi-Gallodoro Unit	Capo Sant'Andrea Unit	

Taormina area				
Somma (1999)		Truillet (1969-70)	Bonardi *et al.* (1976)	Bouillin *et al.* (1992)
Longi-Taormina Unit		Chaine Bordiere	Longi-Taormina Unit	Dorsale Peloritani or Chaine Bordiere
Second-order unit **Taormina** (basement and cover with evolution from continental to neritic platform (Lower Jurassic); from neritic platform to basin (Lower Jurassic); and from basin to bathyal plain (Tithonian-Eocene) Constituted by third-order units	**Embrice 3** basament and cover		C Unit	Costa Sanara slice
	Embrice 2 basament and cover	Taormina Unit		Taormina s.s. Unit
	Embrice 1 basament and cover		B Unit (Costa Ogliastro and M. Ziretto outcrops)	
Second-order unit **Capo Sant'Andrea-Gallodoro** (basement and cover with evolution from continental to neritic platform (lowermost Lower Jurassic); from neritic platform to sea-mount (Lower Jurassic); and from sea-mount to bathyal plain (Tithonian-Eocene)		Longi-Gallodoro Unit	B Unit (Gallodoro and M. Castellaccio outcrops) A Unit	Capo Sant'Andrea Unit

Roccella Valdemone area				
Somma (1999)		Truillet (1969-70)	Bonardi *et al.* (1976)	Bouillin *et al.* (1992)
Longi-Taormina Unit		Chaine Bordiere	Longi-Taormina Unit	Dorsale Peloritani or Chaine Bordiere
Second-order unit **Rocca Licopeti-Pizzo Leo** (basement and cover with evolution from continental to neritic platform (lowermost Lower Jurassic); from platform to slope (uppermost Lower Jurassic); and from slope to shallow marine (Ypresian)		Taormina Unit	B Unit (Roccella Valdemone and Pizzo Leo outcrops)	Rocca Licopeti Unit
Second-order unit **Taormina** (basement and cover with evolution from continental to neritic platform (Lower Jurassic); from neritic platform to basin (Lower Jurassic); and from basin tobathyal plain (Tithonian-Eocene) Constituted by third-order units	**Embrice 2** Lower Jurassic cover			
	Embrice 1 Tithonian - Eocene cover	Longi-Gallodoro Unit	A Unit	

Figure I.2.12. Nomenclature of the subunits within the Longi-Taormina Unit according to several authors, and possible equivalences (translated from Somma, 1999).

In the LTU several subunits, consisting of both Paleozoic basement and Mesozoic-Cenozoic cover, are distinguished, depending on their geometric and geographic location and also on the different sedimentary evolution of the Mesozoic-Cenozoic covers. The main subunits of the LTU have been differently named in the literature (Fig. I.2.12) and therefore a great amount of denominations has been proposed (Duée, 1961; Truillet, 1962, 1969; Caire *et al.*, 1965; Lentini, 1975; Lentini and Vezzani, 1975; Bonardi *et al.*, 1976; Arnone *et al.*, 1979; Bouillin *et al.*, 1992).

Somma *et al.* (2005b) proposed the subdivision of the entire LTU in three subunits, denoted, from base to top, as Lower (Capo Sant'Andrea-Gallodoro), Middle (Taormina) and Upper (San Marco d'Alunzio-Rocca Licopeti) Subunit (Fig. I.2.12). The Lower Subunit (corresponding to the Longi-Gallodoro Unit of Caire, 1960; the Capo Sant'Andrea Unit of Lentini and Vezzani, 1975; the A and B p.p. Units of Bonardi *et al.*, 1976) is exclusively exposed in the eastern outcrops of the LTU, and appears in a few km^2 wide area stretched from Gallodoro to Capo Sant'Andrea (Fig. I.2.8). The Middle Subunit (corresponding to the Longi Nappe of Ogniben, 1960; the Taormina Unit of Lentini and Vezzani, 1975; the B p.p. and C Units of Bonardi *et al.*, 1976; the Longi-Taormina Unit s.s., Lentini and Vezzani, 1975) is the most extensive unit and crops out from the western (Sant'Agata di Militello; Fig. I.2.8) to the eastern outcrops (Taormina; Fig. I.2.8). The Upper Subunit (corresponding to the Galati Nappe of Ogniben, 1960; the S. Marco d'Alunzio Unit of Lentini and Vezzani, 1975; the B p.p. Unit of Bonardi *et al.*, 1976; and the Rocca Licopeti Unit of Bouillin *et al.*, 1992) is exposed in the western and central outcrops, from Sant'Agata di Militello to Rocca Licopeti (Fig. I.2.8).

For the present study, field observations in the Stilo Unit have been made in the Stilo-Bivongi area (Serre Massif) and in the Staiti-Pietrapennata area (Aspromonte Massif), where six and two sections have been measured and sampled for conodonts, respectively. In the Peloritani Mountains six sections belonging to the Longi-Taormina Unit have been measured, sampled and studied for conodonts in the Taormina-Castellmola, Gallodoro, Mongiuffi-Melia, Rocella Valldemone and Pizzo Leo areas. The lithostratigraphic succession in several other outcrops without fossiliferous beds were also checked for reconstruction of a stratigraphic scheme of the Calabrian-Peloritanian Paleozoic.

I.3. Materials and methods

For this dissertation I have used: 1) fieldwork: looking for, studying, measuring and sampling new or previously known Paleozoic outcrops in the selected areas; 2) laboratory work: it was focused on the study of microfacies and, especially, in conodont analysis by i) processing limestone samples in order to extract conodonts, ii) classifying them by study under binocular glass, petrographic microscope and Scanning Electron Microscope (SEM), and iii) making photographs of the best conodontal elements; and 3) deskwork: study of available literature (either regional geological, disciplinar on the Paleozoic stratigraphy and palaeogeography, or biostratigraphical-palaeontological mainly of conodonts), data processing (including picture analysis), discussion of results, drawing, and writing of the papers already published or in preparation, and of this thesis. All sample remains, conodonts found and thin sections are stored in the Departamento de Estratigrafía y Paleontología of the Universidad de Granada (Spain).

I.3.1. Fieldwork

Fieldwork involved 1) exploring the studied areas searching for good outcrops; 2) mapping selected areas; 3) drawing geological sections, paying attention to the outcrop tectonic structure and measuring tectonic microstructures (foliation, folds), when necessary; 4) measuring stratigraphic columns paying special attention to lithology, calcareous and siliceous lithofacies, grain size and sedimentary structures and sequences of structures in siliciclastic lithofacies; and 5) sampling studied sections for conodont and microfacies analyses. In order to carry out microfacies analyses, thin sections of all samples collected have been made by the technicians of the Departamento de Estratigrafía y Paleontología of the Universidad de Granada.

The monotony of the clastic lithofacies, the structural deformation and the spatial discontinuity of the Paleozoic outcrops both in the Betic Cordillera and in the Calabria-Peloritani Terrane make essential to date them, if possible, by means of fossils. This is necessary for correlating different sections and establishing a detailed stratigraphic succession. Conodonts had already been found by previous authors in nearly all the lithostratigraphic units here studied, but mostly in different outcrops (Geel, 1973; Kockel and Stoppel, 1962; de Capoa-Bonardi, 1970; Herbig, 1983; Majesté-Menjoulas *et al.*, 1986; Rodríguez-Cañero, 1993, among others).

I.3.2. Conodont study

It is well known that conodonts are excellent index fossils and endure metamorphic temperatures up to 600ºC. The Paleozoic and Triassic biostratigraphic zonations have been made based, mostly, on the distribution of conodont species because of their rapid evolution and wide distribution (from Cambrian to Triassic). Therefore, a priority of the fieldwork was to look for appropriate horizons for conodont searching, and it has been directed to find limestone beds which may yield conodonts. In addition, and whenever it has been possible, previously published outcrops where others authors had found conodonts have also been sampled and studied.

The size of conodont elements is between a few tens of millimetres and a few millimetres, so they are not usually visible in the field. Consequently, the first sampling was made without productivity information. Whenever it was possible, the amount

sampled was between 1 and 2 kg weight. Sampling intervals have varied depending on the section. A total amount of 523 samples were collected for conodont analysis. Although all of them have been dissolved, 278 of them have been selected to be studied.

I.3.3. Laboratory work

Acetic and formic acids have commonly been used to dissolve carbonate rocks to find conodonts. The protocol used for the dissolution of most limestone samples here studied was that proposed by Jeppsson *et al.* (1999). They demonstrated that the acetic acid protocol traditionally used, published by Graves and Ellison (1941), which used a 10-15% dilute acetic acid solution, destroys most of conodonts and other phosphatic fossils. Therefore, Jeppsson *et al.* (1999) modified this technique by adding calcium acetate as a buffer.

The acetate-buffered acetic acid technique allows to dissolve approximately 60% of the sample prepared per week. On the other hand, the formic acid protocol dissolves almost all the sample in one day. However, during this dissertation I realized that formic acid destroys more conodonts than the acetate-buffered acetic acid technique; so, I decided to minimize their use. Hence, formic acid has been used only when short processing time was crucial.

I.3.3.1. Preparing samples

To promote the dissolution of the carbonates, both with acetic acid and formic acid, is advisable to fragment the rock sample into pieces of 2 or 3 cm in diameter. The standard amount in a preliminary research is usually 1 kg. Nevertheless, in the last times I preferred to prepare 1.2 kg in order to increase the total amount dissolved per time without changing substantially the routine.

• Dissolving samples

In a 12-litre plastic vessel, 10 l of acetate-buffered acetic acid solution is prepared by adding 7.4 l of water, 0.6 l of acetic acid and 2 l of "acetate soup" (see later). In the case of formic acid, the solution is prepared by adding 9 l of hot water and 1 l of formic acid.

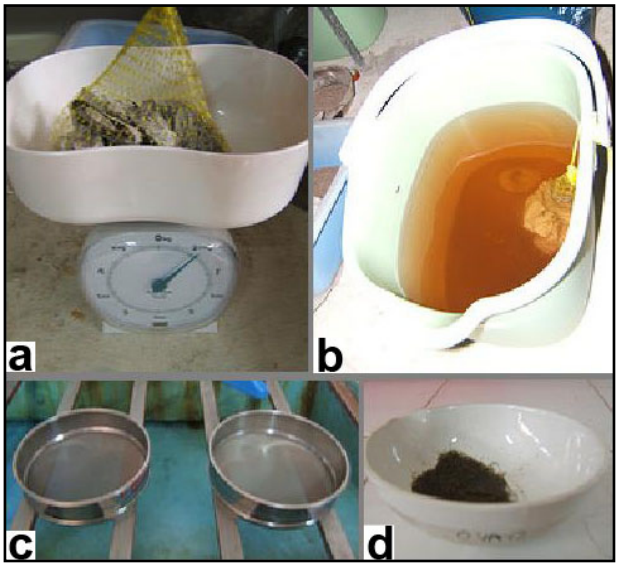


Figure I.3.1. ***a****. Fragmented sample in a plastic net.* ***b****. Plastic vessel and partially dissolved sample.* ***c****. Sieves.* ***d****. Porcelain evaporating dish.*

One kg of fragmented sample is placed in a plastic net (Fig. I.3.1.a) and submerged into the solution (Fig. I.3.1.b). The top of the sample should be very close to the surface of the liquid. The vessel is covered with clingfilm in order to isolate the solution and avoid bad smell.

After one week, it is time to change the solution, and to wash and weight the sample which has not been dissolved. When emptying the vessel, a part of the solution (or acetate soup) is filtered by using filter paper in order to use it as buffer in successive solutions, avoiding stirring the mud in the bottom and left enough acetate soup to continue dissolving the remaining sample after adding enough water and acetic acid to dissolve the rest of the sample. In the case of formic acid solutions, all the sample should be dissolved in one day.

• Sieving

For sieving, I have used two sieves of 160 µm and 56 µm mesh (Fig. I.3.1.c), and recovered only the residue between them. Washing was very carefully made, and the residue placed in a porcelain evaporating dish (Fig. I.3.1.d). It could be air-dried, or within oven.

It is very important to wash properly the sieves

between samples. After using them, they were carefully cleaned with a soft bristle brush. To avoid contamination between samples, the sieves were submerged in methylene blue for at least 2 minutes to dye all the particles in the mesh. Then the sieves were washed with abundant water.

I.3.3.2. Density separation and hand-picking

The residue recovered after the dissolution of limestone samples usually present a low concentration in conodonts. Heavy liquids are advisable to use for density separation when there is too much recovered residue or when the concentration of conodonts is very low. The heavy-liquids density separation method takes advantage of the high specific weight of the phosphatic conodont elements (between 2.84 and 3.10 g/cm^3).

Diverse protocols can be used to remove the banal part in order to facilitate hand-picking. In the first stages of this dissertation, I have used bromoform for density separation. Due to their high toxicity (carcinogenic) we decided to implement the use of sodium polytungstate (SPT) for conodont separations (Anderson *et al*., 1995, and references therein). Both protocols are summarised below.

• Density separation using bromoform

Bromoform is highly toxic and it is necessary to handle it in a fume extractor, wearing gloves and a lab coat. Since all lab glass should be washed with ethanol, and then with abundant water and soap, inside the fume extractor, it must include a sink.

The bromoform is put in a conical graduated decantation funnel and then the residue is added (Fig. I.3.2). Conodonts and heavy minerals fall to the bottom of the funnel because bromoform has lower density (2.89 g/cm^3). The fraction with lower specific weight is suspended at the top. After a few hours, the fraction in the bottom of the funnel is recovered with filter paper, then washed with ethanol and air-dried.

• Density separation using SPT

SPT protocol here used was described by Anderson *et al*. (1995), who improved this technique after years of experimentation using other protocols previously proposed by several authors (Merrill, 1987; Savage, 1988; Morrow and Webster, 1989, among others).

SPT, with a density between 2.76 and 2.79 g/cm^3, is put in a 100 ml polycarbonate test tube and then a low amount of residue. Decantation of heavy metals and conodonts is speeded up using a centrifuge. Then recovered both the light and the heavy fractions using filter papers in a filtering flask attached to a Buchner funnel (Fig. I.3.2). The filter paper is placed on a bench and left to dry at room temperature. Conodonts remain in the heavy fraction; light fraction might contain other microfossils.

• Hand-picking

Once dried, a small amount of the recovered residue is scattered on a metallic tray (aprox. 10 x 6 cm) with squared background. Each square is examined using a binocular microscope and the conodont elements found recovered with a very thin moistened brush and stored into a microslide. A non-greasy gum has been placed in the microslides to facilitate their order and study under magnifying glass and/or microscope.

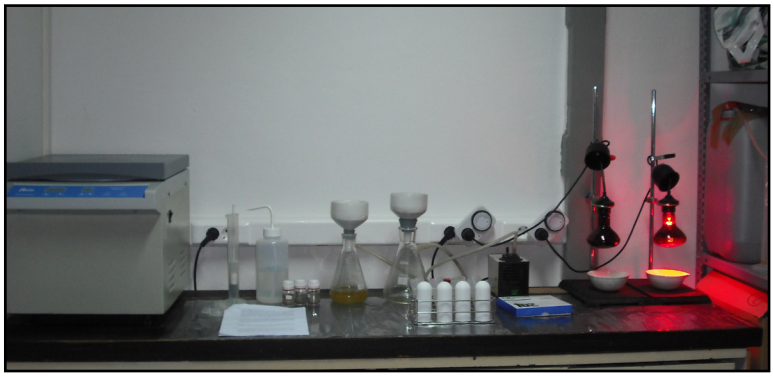


Figure I.3.2. Density separation using bromoform (left) and sodium polytungstate (right).

I.3.3.3. Electron microscopy study

The most representative and best preserved specimens have been photographed using scanning electron microscopy in the Centro de Instrumentación Científica of the Universidad de Granada. Secondary electron images were performed on both carbon and gold-coated samples with ESEM (FEI, mod. Quanta 400, Environmental Scanning Electron Microscopy), VPSEM (LEO 1430-VP Variable Pressure Scanning Electron Microscope) and SEM (Zeiss DMS Scanning Electron Microscope). Working voltage was set at 10-15 kV.

I.3.4. Online storage data: www.conodonta.org

The project conodonta.org is a web application for the management of rock samples that have been designed for this reseach, in order to easily storage section and sample data. It is free software with GPL license and has been developed in PHP scripting language using a MYSQL database with five related tables. Although it has been developed to storage information about conodont samples, it can be easily adapted for working with other kinds of rock samples. It also could be extended for the geolocation of the sections and samples using GIS software.

The interface shows five areas: sections, samples, dissolutions, conodonts and images, where we can look up, introduce and modify fields, like name section, sample stock, location in the storage room or information about the dissolutions, as well as taxonomic and biostratigraphic data and other observations.

Online access and posibility of exporting the information easily to spreadsheets, xml files, etc., prevents data lost and facilitates work.

CHAPTER II

Paleozoic of the Malaguide Complex

II. 1. The Vélez Rubio - Lorca Corridor
II. 2. The Cogollos Vega Zone
II. 3. The Ardales area
II. 4. Synthesis of the Malaguide Paleozoic stratigraphy

II.1.1. Lithostratigraphic features of the studied tectonic units

II.1.1.1. Geological setting and previous regional studies

The Vélez Rubio-Lorca Corridor (Fig. II.1.1) is a ENE-WSW narrow belt along the tectonic contact between the Internal and the External zones of the Betic Cordillera, in the north of the province of Almería. This belt connects the post-orogenic basins of Guadix-Baza (Granada province) and Lorca (Murcia province). The Vélez Rubio-Lorca Corridor is occupied in a great extend by Quaternary alluvial sediments that result from the erosion of both the Subbetic Complex, to the north (Sierras de Orce and Maria), and the Alpujarride Complex, to the south (Sierra de las Estancias). Under these recent sediments appears one of the most important Malaguide outcrops. In particular, in the eastern sector of the Betic Cordillera, here is where Paleozoic Malaguide rocks mostly crop out. In the other important outcrop of this complex in this sector, Sierra Espuña (Murcia province), Mesozoic-Cenozoic (Triassic-lower Miocene) terrains predominate (Martín-Martín, 1996, and references therein).

Along the Vélez Rubio-Lorca Corridor, to the north, the Malaguide Complex is tectonically overlied by detritic sediments of the Campo de Gibraltar Complex, which is in turn overthrusted by the Subbetic Complex (Fig. II.1.1). To the south, the Paleozoic rocks of the Malaguide Complex cover the Triassic phyllites and carbonates of the Alpujarride Complex (Fig. II.1.1) along a N-dipping surface that constitutes a major extensional tectonic contact (Lonergan and Platt, 1995; Jabaloy *et al*., 2007, among many others).

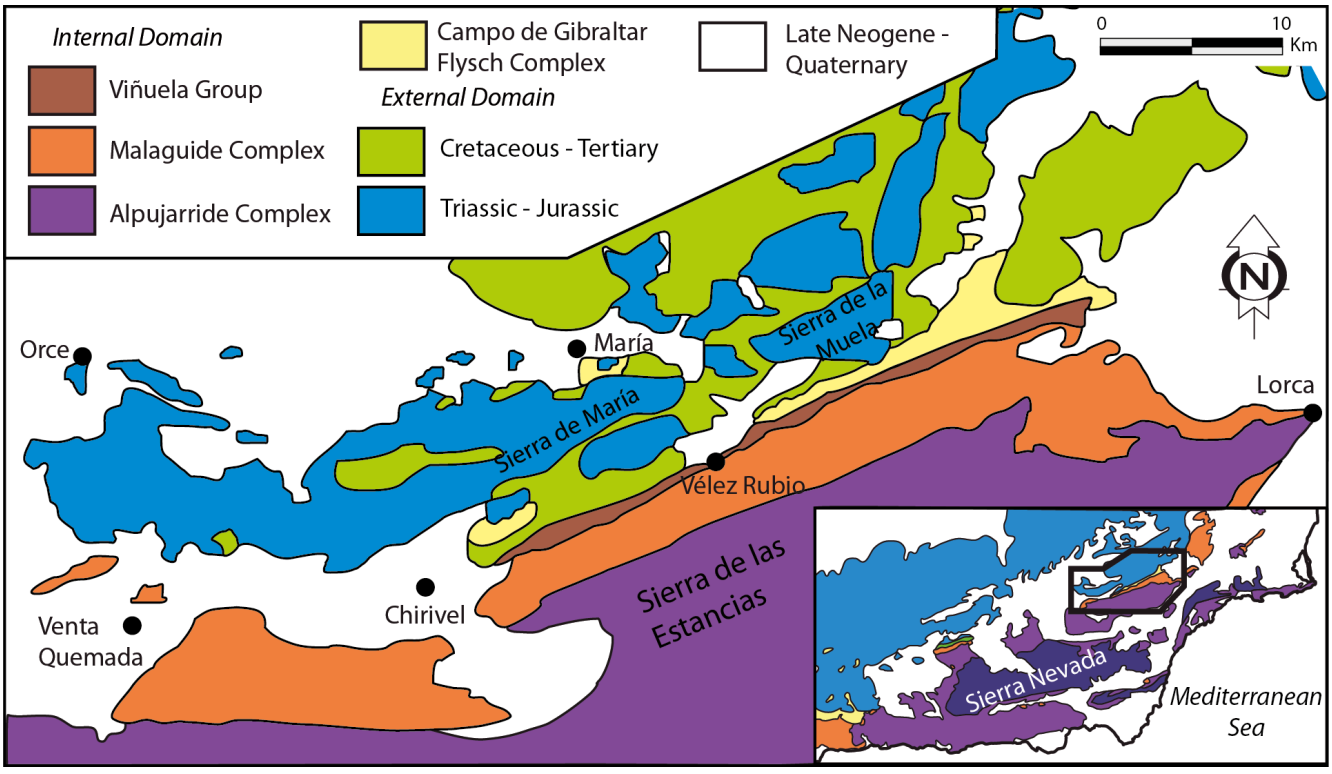


Figure II.1.1. Tectonic sketch of the Vélez Rubio-Lorca Corridor (modified from Martín-Algarra, 1987).

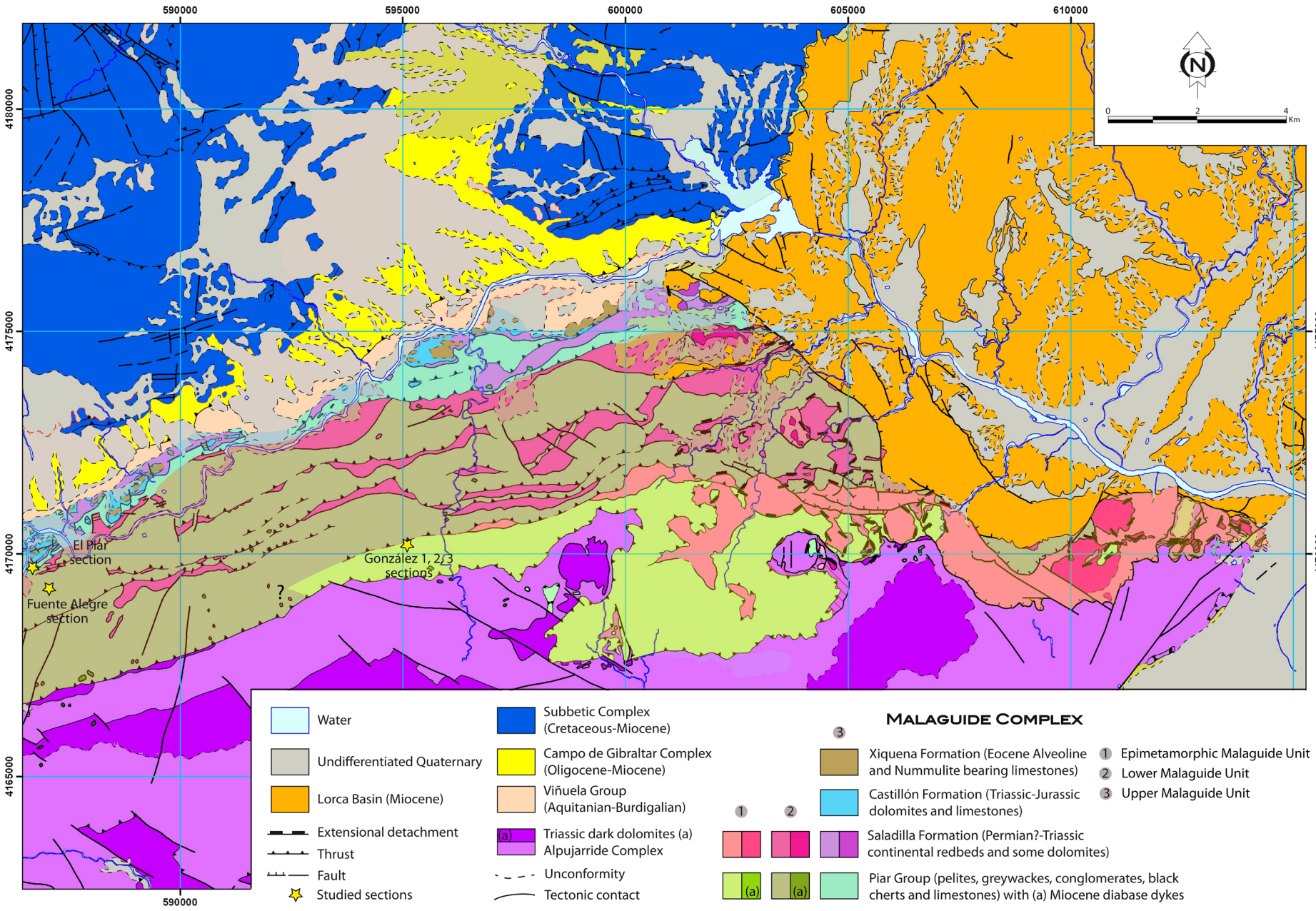


Figure II.1.2. Malaguide Alpine units distribution in the eastern Vélez Rubio-Lorca Corridor with location of the studied sections (modified from Plan Magna maps 952, 953, 974, 975: Baena Pérez et al., 1977; Kampschuur et al., 1981; Voermans et al., 1979; Martín Vivaldi et al., 1974, respectively).

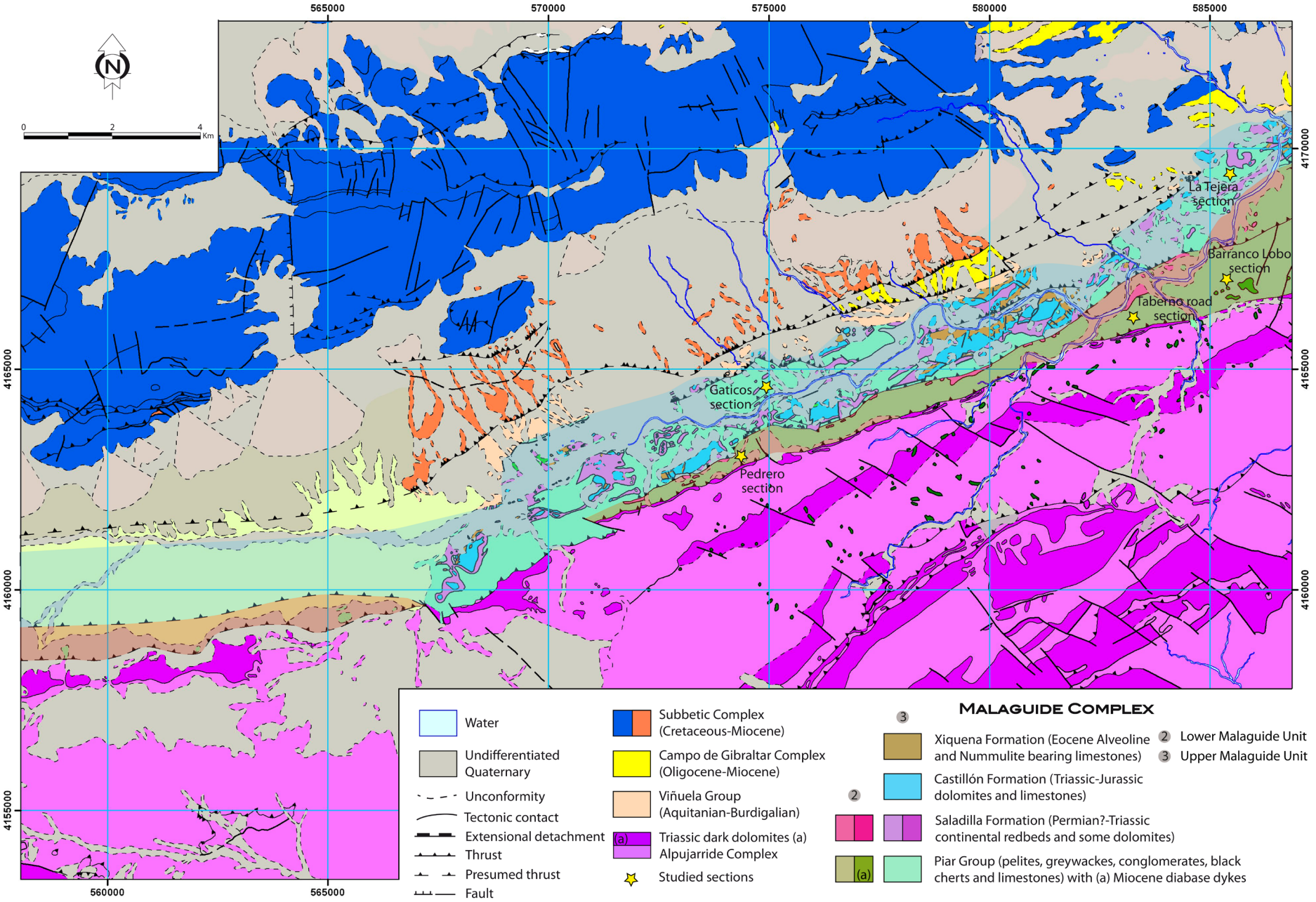


Figure II.1.3. Malaguide Alpine units distribution in the central area of the Vélez Rubio-Lorca Corridor with location of the studied sections (modified from Plan Magna maps 951, 952, 973, 974: Baena Pérez et al., 1977, 1979a, 1979b; Voermans et al., 1979, respectively).

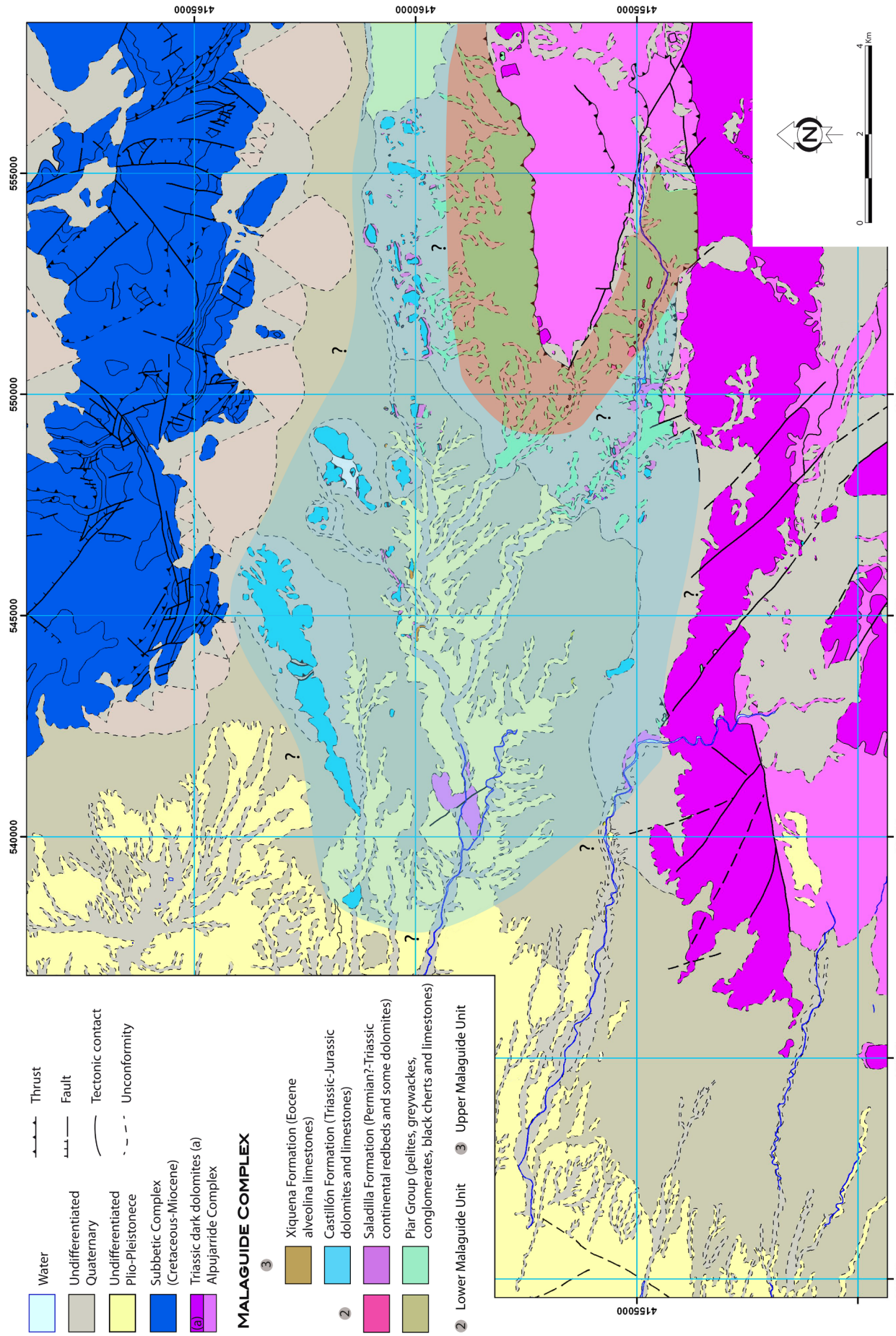


Figure II.1.4. Malaguide Alpine units distribution in the western Vélez Rubio-Lorca Corridor with location of the studied sections (modified from Plan Magna maps 973: Baena Pérez et al., 1979a).

The Malaguide Complex of the Vélez Rubio-Lorca Corridor was studied in detail by several Dutch geologists (Mac Gillavry *et al*., 1960, 1963; Roep and Mac Gillavry, 1962; Geel, 1967, 1973; Soediono, 1970; Roep, 1972; Mäkel, 1995). These authors recognized, and actually defined for the first time, the main lithostratigraphic units that nowadays are differentiated in this complex. They assigned Paleozoic terrains to the Piar Formation, later named Piar Group by Martín-Algarra (1987); predominantly continental Triassic deposits to the Saladilla Formation; they also differentiated several lithostratigraphic units in the Malaguide Mesozoic-Paleogene dolomites and limestones, and defined the Ciudad Granada Group.

These authors pointed out the high deformation that affects the Malaguide Complex in this area, where Roep and Mac Gillavry (1962) distinguised four Alpine tectonic units with different stratigraphic successions. All these units show lateral continuity along almost all the corridor and are separated by N-dipping tectonic contacts. They are subdivided in slices with limited lateral continuity. From bottom to top tectonically, and from south to north geographically, these units were named Casolidad, Cerro Colorado, Castillos and Salud (figs. II.1.2, II.1.3 and II.1.4).

Field observations made in this area during this dissertation confirm the broad validity of this division of the Malaguide Complex in units and slices (see also Plan Magma maps, sheets 952, 953, 973, 974, 975), although regarding to the Paleozoic successions, some clarifications are needed. Alpine tectonic units with similar features are also recognized in other sectors where the Malaguide Complex crops out along the boundary between the Internal and External Zones, both in the Granada province (Cogollos Vega Zone of Blumenthal and Fallot, 1935; see section II.2) and in diverse areas in the Málaga province (Serranía de Ronda, Casares-Gaucín and Ardales; see section II.3). These Alpine Malaguide units, from here forth, will be named (figs. II.1.2, II.1.3 and II.1.4): Epimetamorphic Malaguide Unit (the lowermost one), Lower Malaguide Unit (equivalent to the Cerro Colorado Unit), and Upper Malaguide Unit (that includes both Castillos and Salud units, which do not show differences between their Paleozoic successions).

II.1.1.2. Previous lithostratigraphic studies on the Malaguide Paleozoic

Exhaustive studies related to the Paleozoic stratigraphy of the Malaguide Complex in the Vélez Rubio-Lorca Corridor have been made only by Soediono (1970) and Geel (1973). They proposed slightly different stratigraphic divisions that they correlated to the Malaguide Paleozoic differentiated in the *Montes de Málaga*.

Soediono (1970) differentiated the following members stratigraphically downwards (Fig. II.1.5):

• *Unit D*: Upper clastic unit. It is made of very coarse dark greyish-green, usually clast-supported conglomerate with sandy matrix in variable amount. The components of this conglomerate are cobbles and boulders of dark-grey fossiliferous limestones, pebbles and cobbles of gneisses (radiometric age in one sample: 535 ± 75 Ma), black cherts, quartz grains and olive-green sandstones.

• *Unit C*: Dark limestone unit. It mainly consists of thin-bedded, dark-grey to black-coloured limestones. This limestone unit frequently grades into platy shales, and occasionally it intercalates beds of dark coloured pelite and sandstone. As in the other units, tectonic deformation is strong and results in contorted beds and strongly brecciated rocks.

• *Unit B*: Lower clastic unit. It comprises, mainly, greywackes and pelites with typical brownish-green weathering colours, and also conglomerates. Few intercalations of dark thin-bedded limestones are also included. The conglomerates are thick- to very thick-bedded, and are mainly composed by well-rounded granules and pebbles of dark coloured sandstone, white quartz and black chert, which are included in a greyish-brown matrix. The greywackes are thin- to very thick-bedded sandstones with characteristic olive-green colour and still preserve some sedimentary structures (sole markings, grading and slumping). The very thin-bedded olive-green pelites commonly grade into greywackes without a detectable break, and show frequent current lamination.

• *Unit A*: Chert, fossiliferous limestones and variegated pelites unit. This unit is composed of four types of lithofacies:

1. The main lithology is a thinly laminated to thin-bedded, dark-grey to black chert. This chert is strongly brecciated and intruded by light coloured quartz veins. The total thickness is usually less than 5 m but occasionally may reach 35 m.

2. Dark-grey, thin-bedded limestones and olive-green, very thin-bedded tentaculites-bearing limestones. These limestones are also strongly brecciated.

3. A buff-coloured, strongly brecciated, very well indurated, massive limestone with Orthoceras and Silurian conodonts (van den Boogaard, 1965) as well as Silurian tintinnids (Hermes, 1966). This lithology was found only in one locality as one isolated block, and it is not associated with the other rocks of his *Unit A*. Soediono (1970, pag. 32) decided to include this rock type in this unit only for paleontological and time-stratigraphic considerations. Unfortunately, I was not able to find this block in the field.

4. Reddish and greenish thin-bedded clastic rocks associated with the cherts were also included in this unit.

On the other hand, Geel (1973) distinguished, from top to bottom, the following members (figs. II.1.6 and II.1.7):

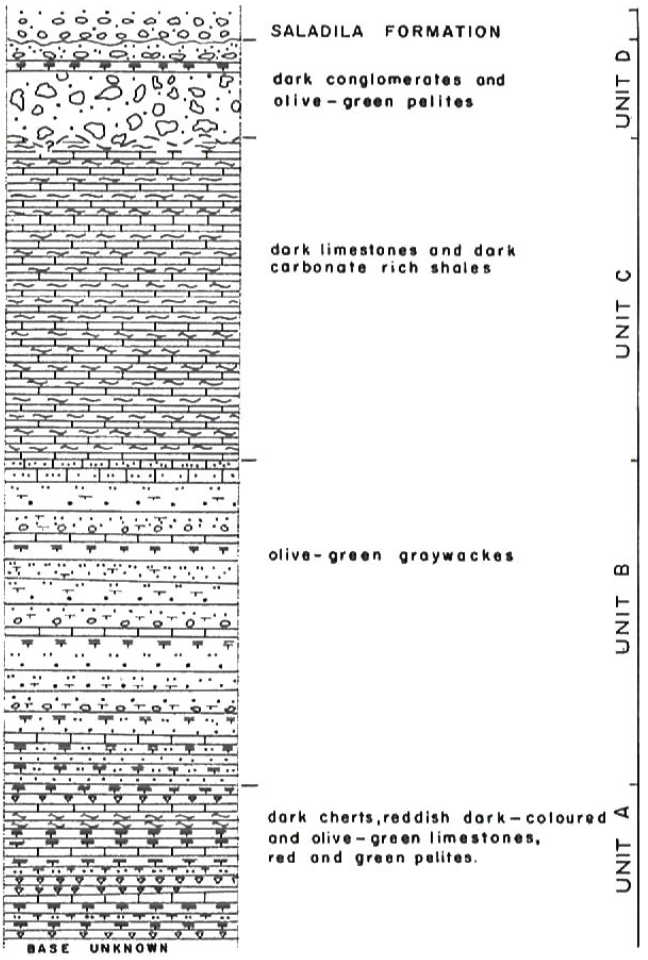


Figure II.1.5. Synthetic stratigraphic column of the Malaguide Paleozoic in the Chirivel area according to Soediono (1970).

• *Marbella conglomerate member*: it consists of very thick conglomerate beds. The main components of the conglomerate are: greywackes, siltstones, shales, sandstones, cherts, granites, gneisses, phyllites and limestones. The latter yielded Visean foraminifera. Except for the limestone boulders, which may reach 20 m, the maximum grain size is 12 cm.

• *Polymict conglomerate member*: it forms coarser intervals within the *Greywacke-shale member*, and grade laterally into conglomeratic greywackes. The main components of the conglomerate are sandstones, greywackes, quartzites, cherts, granites, gneisses and limestones. The latter yielded Famennian and Early Carboniferous conodonts.

• *Greywacke-shale member*: it comprises finer and coarser variants. The finer ones are composed of finely laminated to very thin-bedded bluish shales, and thin-bedded to medium-bedded bluish, greenish and brownish quartzitic siltstones with intercalations of dark laminated to very thin-bedded cherts and thin-bedded to medium-bedded limestone. The coarser variants consist of thin-bedded to very thick-bedded greywackes, bluish-grey when fresh, green or reddish-brown when weathered. The greywackes sometimes show complete Bouma sequences or base cut-out sequences and few sole markings. This member constitutes most of the Paleozoic outcrops in the area studied by Geel (1973).

• *Calizas alabeadas member*: it is made of very thin-bedded to thick-bedded, dark blue or brownish limestones, calcareous sandstones and siltstones, with intercalations of laminated to medium-bedded yellow and brown greywackes and siltstones, and finely laminated olive-green shales. Complete Bouma sequences are sometimes visible in the calcareous sandstones and greywackes, but truncated base cut-out sequences are more common, as well as sole markings and other sedimentary structures.

• *Conodont limestone member*: it consists of very thin-bedded blue limestones, laminated to very thin-bedded bluish and brownish slaty shales,

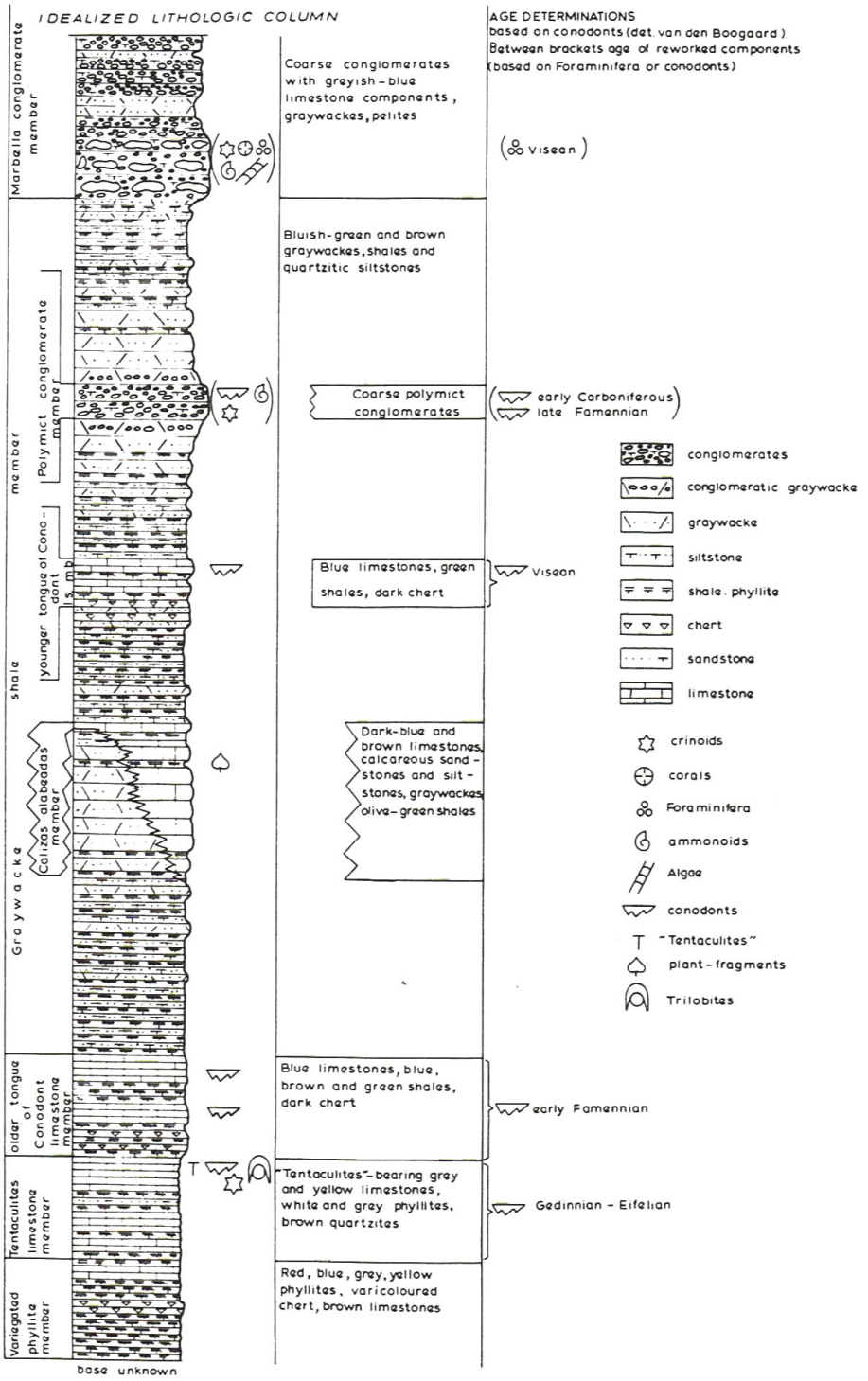


Figure II.1.6. Synthetic stratigraphic column of the Malaguide Paleozoic in the Vélez Rubio area according to Geel (1973).

and laminated to thin bedded dark chert. Because of the conodont fauna found in two samples from diverse localities, Geel (1973) differentiated between an *older tongue* (early Famennian, Upper Devonian) and a *younger tongue* (Visean, Lower Carboniferous) within the member.

• *Tentaculites limestone member*: it comprises finely laminated to thin-bedded greyish and yellowish limestones with their surface crowded with tentaculites (dacryoconarids). White and greyish shales with brown quartzites are locally intercalated. Besides tentaculites, the limestones also yield Early Devonian conodonts.

• *Variegated phyllite member*: it is made of reddish, bluish, greyish and yellowish, very thin laminated to thin-bedded phyllites, and black, locally varicoloured, laminated to thin bedded chert. Occasionally, these lithotypes alternate with laminated to thin-bedded limestones.

Within their very detailed stratigraphic frame, Soediono (1970) and Geel (1973) were able to distinguish the main lithological units present in the Malaguide Paleozoic of their study areas. However, and because of the tectonic complexity, the scarcity of palaeontological data and the lack of knowledge about the Paleozoic stratigraphy of the Malaguide Complex at that time, these authors were not able to propose a uniform stratigraphic scheme. Moreover, according to the present state of knowledge of the Malaguide stratigraphy, several of their proposed stratigraphic relationships between their members/units can be considered as not fully correct, or even wrong. Actually, in spite of working contemporaneously in adjacent areas, they did not make a unique proposal for the subdivision of their Piar Fm. Nonetheles, some of their units are completely equivalent and can be strictly correlated (see also Geel, 1973). In particular:

• Soediono's *Unit D* is similar to *Marbella conglomerate member* of Geel (1973), and both lithological units are totally equivalent to the Marbella Fm.

• The strongly folded dark limestones and dark carbonate rich shales of the Soediono's *Unit C* are analogous to the *Calizas alabeadas member* of Geel. Both are comparable to the Santi Petri Fm.

• In his *Unit B*, Soediono included chert pebbles conglomerates bearing, which probably are analogous to those of the *Polymict conglomerate member* of Geel. On account of the conglomerate components, grain size and matrix nature described by both authors, these units are probably equivalent to the Retamares conglomerate (Almogía Fm.).

• Soediono also included in his *Unit B* olive-green greywackes and pelites comparable to the *Greywacke-shale member* of Geel. Most of these two later units are equivalent to the finer facies of the Almogía Fm., although some greywacke and pelite outcrops, which they included together, belong, in fact, to the Morales Fm.

• The cherts in the Soediono's *Unit A* and those of the *Variegated phyllite member* of Geel should be the same key horizon dated as Tournaisian (Lower Carboniferous) in the western outcrops of the Malaguide Complex (O'Dogherty *et al*., 2000), and that constitutes the lower member of the Falcoña Fm.

• Both Soediono and Geel described tentaculites bearing limestones, although they placed them in different stratigraphic positions.

• The Silurian orthoceras limestones in the *Unit A* of Soediono were not found by Geel in her study area. This lithological unit along with the tentaculites bearing limestones probably corresponds to a different formation.

• Geel defined a *Conodont limestone member* that Soediono did not found in his area. As mentioned before, she distinguished a *younger* and an *older tongue*. Because of the *younger tongue* yielded Visean conodonts, their assignment to the upper member of the Falcoña Fm. is unequivocal. The *older tongue* should belong to a different formation, which, as well as the tentaculites bearing limestones, occurs in a stratigraphic position equivalent to the Santi Petri Fm.

II.1.1.3. Epimetamorphic Malaguide Unit

The Casolidad Unit is the Epimetamorphic, and lowermost, Malaguide Unit of the Vélez Rubio-Lorca Corridor. This unit was defined by Roep and Mac Gillavry (1962) to the south of La Parroquia village (Murcia province) and named after a group of farmhouses to the ESE of the Cerro del Piar (Fig. II.1.2). Roep (1972) pointed out that this unit continues to the east in the Peña Rubia Unit and is affected by an Alpine metamorphic deformation and recrystallization more intense than in other Malaguide units. In fact, because of being metamorphic, similar terrains in other areas have been assigned to the Alpujarride Complex, whereas Sanz de Galdeano *et al.* (1995a, b, 1999) included them in the Transitional units between the Malaguide and the Alpujarride complexes.

The Paleozoic rocks of the Epimetamorphic Malaguide Unit (Fig. II.1.2) tectonically overlie Triassic grey-bluish phyllites of the Sierra de la Torrecilla (Alpujarride Complex, W Lorca). The Triassic Saladilla Fm. of this Malaguide Unit is overlied by the Paleozoic terrains of the Lower Malaguide Unit by means of a tectonic contact that disappears 2 km to the west of the Cortijo de los González, between pelites of the Morales Fm. (Fig.

II.1.2).

Alpine metamorphism in the Epimetamorphic Malaguide Unit is more clearly observable in Triassic rocks (transformed into silky pelites and red phyllites with pinkish and whitish quartzites) than in the Paleozoic Malaguide terrains. These Triassic rocks widely crop out in the area next to Lorca but, to the west, they narrow and finally disappear, making almost impossible to distinguish Paleozoic rocks from the Epimetamorphic and Lower Malaguide units. Despite they are slightly metamorphic, these Triassic rocks are totally similar to those of the Saladilla Fm. Actually, they were included by Roep (1972) in this formation together with the Peña Rubia of Lorca.

Dark terrains constituted mainly by metapelites and metapsammites with microconglomerate beds that can be assigned to the Morales Fm., appear underlying the Triassic Epimetamorphic Malaguide. Locally, to the east, they also include calcareous turbidites (*Calizas alabeadas* facies) that can be ascribed to the Santi Petri Fm. These Paleozoic terrains are cut by diabase dykes.

In this Epimetamorphic Malaguide Unit, I have made field observations along the Rambla de los Arcos, Rambla Saceja and la Alquería, in the surroundings of the Ermita de los Morotes, and mostly along the road from Venta de las Macetas to Lorca and Fuensanta-La Parroquia, particularly in the surroundings of the Cortijo de los González, as detailed below.

II.1.1.4. Lower Malaguide Unit

This unit was named Cerro Colorado Unit by Roep and MacGillavry (1962). The Lower Malaguide Unit is the widest unit of the Malaguide Complex in the Vélez Rubio Corridor and crops out from Venta Quemada to the Lorca Basin. This unit directly overlies the Alpujarride Complex where the Epimetamorphic Malaguide Unit is absent (figs. II.1.2, 3, 4 and 7). The Paleozoic succession of the Lower Malaguide Unit is characterised by four of the five formations classically differentiated in the Piar Group: Morales, Santi Petri, Falcoña and Almogía. Regarding to the facies, this unit is mostly characterised by the presence of the typical *Calizas alabeadas* of the Santi Petri Fm.

Permian?-Triassic terrains of the Lower Malaguide Unit are made up of sandstones and pelites in Pseudoverrucano facies (not affected by metamorphism), which includes Triassic dolomites thinner than in the Epimetamorphic Malaguide Unit, and gypsum levels (Geel, 1973). Jurassic and younger formations only appear sporadically (figs. II.1.3, 4 and 7).

In the Lower Malaguide Unit, I have made numerous field observations and measured sections in the following areas (figs. II.1.3, 4 and 7): to the south of Venta Quemada, Barranco Pedrero (SW of the Cerro de la Monja), road to Taberno (SE of Vélez Rubio), along the Barranco Lobo and in the Umbría de Fuente Alegre (Barranco Hondo and Barranco de Fuente Alegre).

II.1.1.5. Upper Malaguide Unit

Roep and Mac Gillavry (1962) distinguished two tectonic units above the Cerro Colorado Unit: Castillos (after the castles of Vélez Rubio and Xiquena) and Salud (after the Ermita de la Salud in Vélez Rubio) that differ in the features of their Mesozoic and Cenozoic successions.

The Castillos Unit include widest, better and stratigraphically more complete outcrops of Jurassic, Cretaceous and Paleogene Malaguide formations. The Salud Unit occupies a minimum surface and is mainly characterised by including the Ciudad Granada Group. Concerning their Paleozoic formations, there are no differences between them. Because of that, from here on, both units will be consider minor subdivisions of only one unit, the Upper Malaguide Unit (figs. II.1.2, 3, 4 and 7).

Paleozoic terrains of the Upper Malaguide Unit are different from those of the other two Malaguide units (Epimetamorphic and Lower Malaguide units). The Paleozoic succession of the Upper Malaguide Unit does not show neither the Morales Fm. nor the Santi Petri Fm., whereas, on the contrary, its

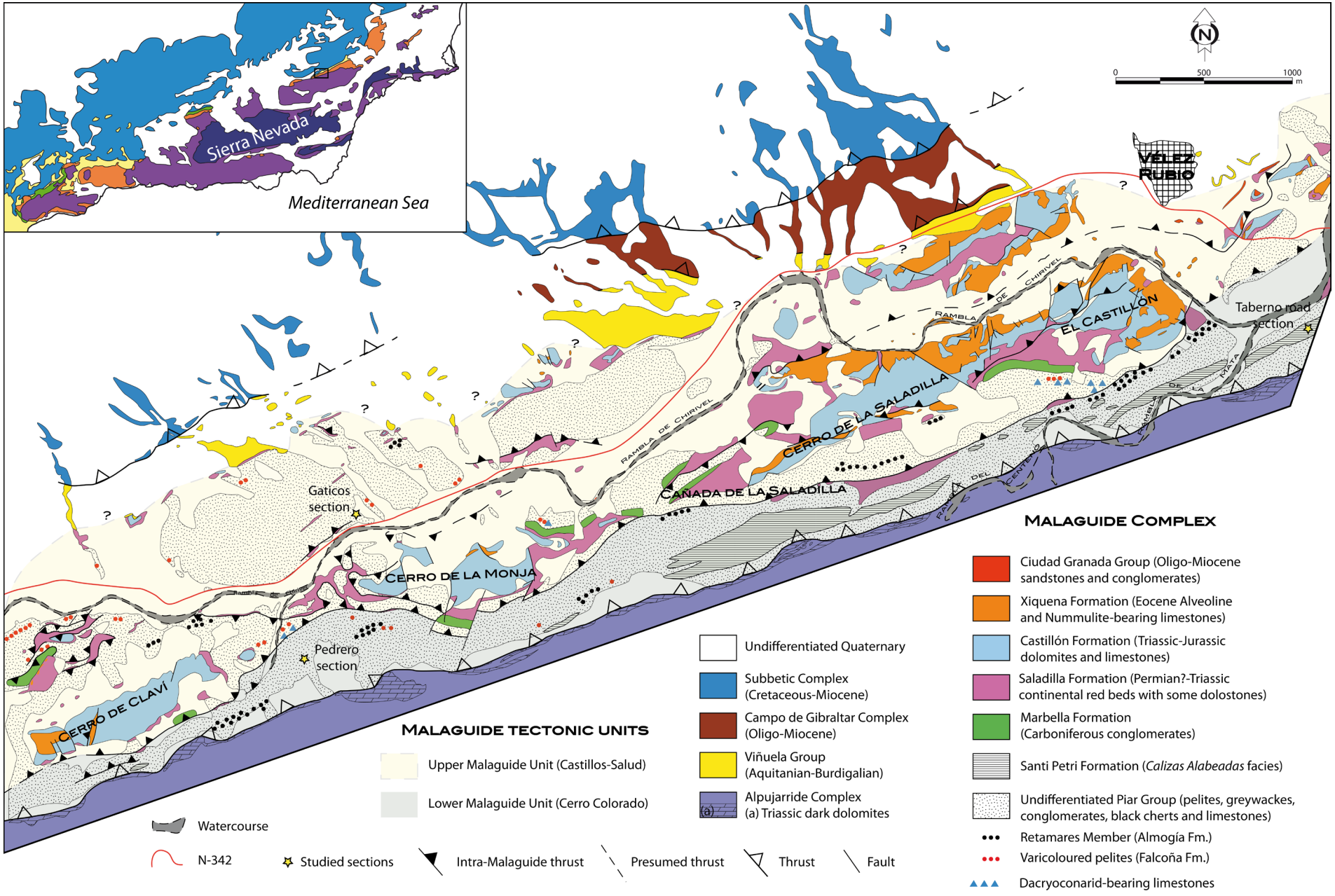


Figura II.1.7. Geological map of the Malaguide formations and Alpine units to the west of Vélez Rubio and location of the studied sections (modified from Geel, 1973).

stratigraphic succession includes thin limestone lenses that yield Silurian and Devonian microfossils (dacryoconarids, tintinnids and conodonts). Such levels were included by Soediono (1970) in his *Unit A* (lithological groups 2, 3 and probably 4, Fig. II.1.5), and by Geel (1973) in her *Variegated phyllite member*, *Tentaculites limestone member* and *Older tongue* of the *Conodont limestone member* (Fig. II.1.6). These limestone intercalations make up thin levels (cm- to m- thick), included in slates or calcareous slates with a deformation and metamorphic recrystallization much lower than in the Santi Petri and Morales formations. Such slates with thin limestone levels include few sandstone intercalations that sometimes show varicoloured tonalities, and for that reason Geel (1973) included them in her *Variegated phyllite member*, that she also recognised, however, in her Cerro Colorado Unit (Fig. II.1.7).

Directly related to these Paleozoic terrains, in many outcrops appear thin-bedded radiolarites of the Falcoña Fm. (up to 30 m thick, p.e. to the north of the A-92 highway, in the Barranco de Reolí, Barranco de las Peñicas, along the Rambla de Chirivel close to the Rambla del Claví, Fig. II.1.5). Such rocks are frequently related to Visean limestones, but until now, and in the Upper Malaguide Unit in the Vélez Rubio-Lorca Corridor, they do not released any conodont remain. Therefore, their assignment to the Falcoña Fm. could not be totally demonstrated. In any case, lydites found in the Upper Malaguide Unit seem to be less deformed and recrystallized than in the comparable levels of the Falcoña Fm. observed in the Lower Malaguide Unit.

Most of the Paleozoic outcrops of the Upper Malaguide Unit in the area between Chirivel, Vélez Rubio and La Parroquia correspond to slates and greywackes assignable to the Almogía Fm., with several isolated outcrops of conglomerates facies of Retamares type.

The main outcrops of the Marbella Fm., ascribed by Soediono (1970) to his *Unit D,* and by Geel (1973) to her *Marbella conglomerate member*, are located in the Upper Malaguide Unit, in particular, in the Castillos Unit, where Geel (1973) mapped it in numerous outcrops (Fig. II.1.7).

In the Upper Malaguide Unit I have studied, measured and sampled sections mainly in the Vélez Rubio area (Fig. II.1.2, 3 and 7). In the area to the south of Vélez Rubio: Barranco Reolí (to the NW of the Cerro de la Monja) and to the south of El Castellón; to the south of Cerro de la Tejera and to the south of the Cerro del Piar; and in the eastern end of this Malaguide Unit, next to the Lorca Basin, in the surroundings of the Venta del Coronel.

II.1.2. Previous biostratigraphic data

Silurian, Devonian and Lower Carboniferous rocks have been previously dated in the Vélez Rubio area. The first published dating in this area was made by Roep (Mac Gillavry *et al.*, 1960), who described the occurrence of *Novakia* sp. indet. within limestone lenses in the Xiquena castle area. As the dacryoconarids could not be specifically identifiable, only a Devonian age was inferred.

van den Boogaard (1965) published two conodont faunas extracted from limestone lenses, of late Silurian and earliest Famennian age. The first sample came from a site located to the SE of Chirivel, and was collected by Soediono from a metric bed of reddish, brecciated limestone with orthoceras. This sample yielded *Kockelella variabilis* Walliser, 1957, among other conodonts, which indicates a late Silurian age. They did not located this sample in either of the tectonic units mentioned in the previous section, but according to its provenance, this sample was collected from one outcrop belonging to any of the above mentioned upper Malaguide units, probably from the Castillos Unit. Soediono (1971) included the sampled outcrop in his lithological *Unit A*. One year latter, Hermes (1966) dated exactly the same limestone in the same outcrop by means of tintinnids.

The other sample studied by van den Boogaard (1965) was collected from one lens of grey limestone in the Cerro de La Tejera, and yielded a conodont fauna association including *Palmatolepis quadrantinodosalobata* Sannemann, 1955a among others, which indicates a lowermost Famennian age (Upper Devonian).

Geel (1973) mentioned the presence of Visean conodonts to the SW of the Cerro de la Monja but no more details were provided. At that site, and also in the Fuente Alegre area, we have reported in two preliminary papers of this thesis, several conodont limestone beds of Visean age and closely associated

with thin bedded radiolarian cherts, which are correlated to the two members of the Falcoña Fm. in the Málaga area (Navas-Parejo *et al*., 2006, 2008, see appendix I).

Finally, Geel (1973) also found Visean foraminifera and conodonts, and other fossils without biostratigraphic information, in limestone pebbles from her *Marbella Conglomerate member* (equivalent to the Marbella Fm.) to the south and west of the Cerro del Castellón of Vélez Rubio. Later, Buchroithner *et al*. (1980) studied in detail the microfacies and conodont fauna of the Retamares member (Almogía Fm.) and the Marbella Fm., and pointed out a Givetian, Upper Devonian and Lower Carboniferous age to the limestones pebbles of the Retamares member and a Late Visean-Bashkirian age to the limestone pebbles and boulders of the Marbella Fm.

II.1.3. Studied sections

In addition to the many local observations made when looking for outcrops to sample for conodonts, several outcrops were selected for a more detailed study, in order to reconstruct the stratigraphic succession of the Malaguide Complex units in the Vélez Rubio-Lorca Corridor. In spite of strong tectonic deformation in all of them, it is yet possible to identify the most typical facies of the several Malaguide Paleozoic formations, to recognize at least some of their sedimentary features and stratigraphic relationships, sometimes to date them by conodonts, and to reconstruct the main features of the general stratigraphic column in different units and areas. The location of these outcrops is indicated in figs. II.1.2 and II.1.5, and a brief outline of them is given in the following, from those found in the Epimetamorphic, Lowermost Malaguide Unit to those of the Lower Malaguide Unit to, finally, those of the Upper Malaguide Unit.

II.1.3.1. Epimetamorphic Malaguide Unit

The most interesting stratigraphic observations in this unit have been made along the road from Henares to La Parroquia, in particular in the area around the Cortijo de los González (figs. II.1.2 and II.1.8), where three sections with Devonian and Carboniferous beds have been reported, and dated (sections 1, 2 and 3 in figs. II.1.8 and II.1.9; for biostratigraphic details, see paragraph II.1.4.).

The Cortijo de los González is reached from the A-91 highway, exit 6, to the north about 10 km along an unnamed road towards La Parroquia village, road MU-701. Unfortunately, the González 3 section (located just to the west of the Cortijo de los González, which was discovered and partially studied and sampled in 2005 and 2006) has been destroyed by the installation of a gas pipeline made in 2008 (Fig. II.1.10, see also Fig. II.1.8.C).

North of the Venta de las Macetas and along the road from Henares to Fuensanta-La Parroquia, the silvery grey and pinkish phyllites of the Alpujarride Triassic and the overlying brownish, low-grade metamorphic, Paleozoic slates belonging to this Epimetamorphic Malaguide Unit, form a succession gently dipping N. These Paleozoic rocks are crosscut by metre- to (rarely) decametre-thick dykes of basic subvolcanic rocks, and are identical to the rocks belonging to the Morales Fm. in the Málaga area. Nonetheless, they also remarkably remind the lowest grade brown schists from the Alpujarride Complex that can be observed underlying the Triassic phyllites along the same road as well as along the A-92N highway and surrounding areas.

About 300 m to the south of the Cortijo de los González (Fig. II.1.8.A and B), the contact with the Alpujarride rocks is well exposed and the succession starts with alternating thin to medium-bedded brownish-greyish slates and thin and fine-grained metasandstone beds. There, they also intercalate a several tens of m-thick interval including black, thin-bedded and fine-grained detrital limestones very similar to the *Calizas alabeadas* facies (but less folded at that site) that make most of the outcrops of the Santi Petri Fm., which is widespread in the overlying Lower Malaguide Unit. The contact between the latter interval and the beds outcropping at the Cortijo de los González is not clearly visible, but the same rock succession of

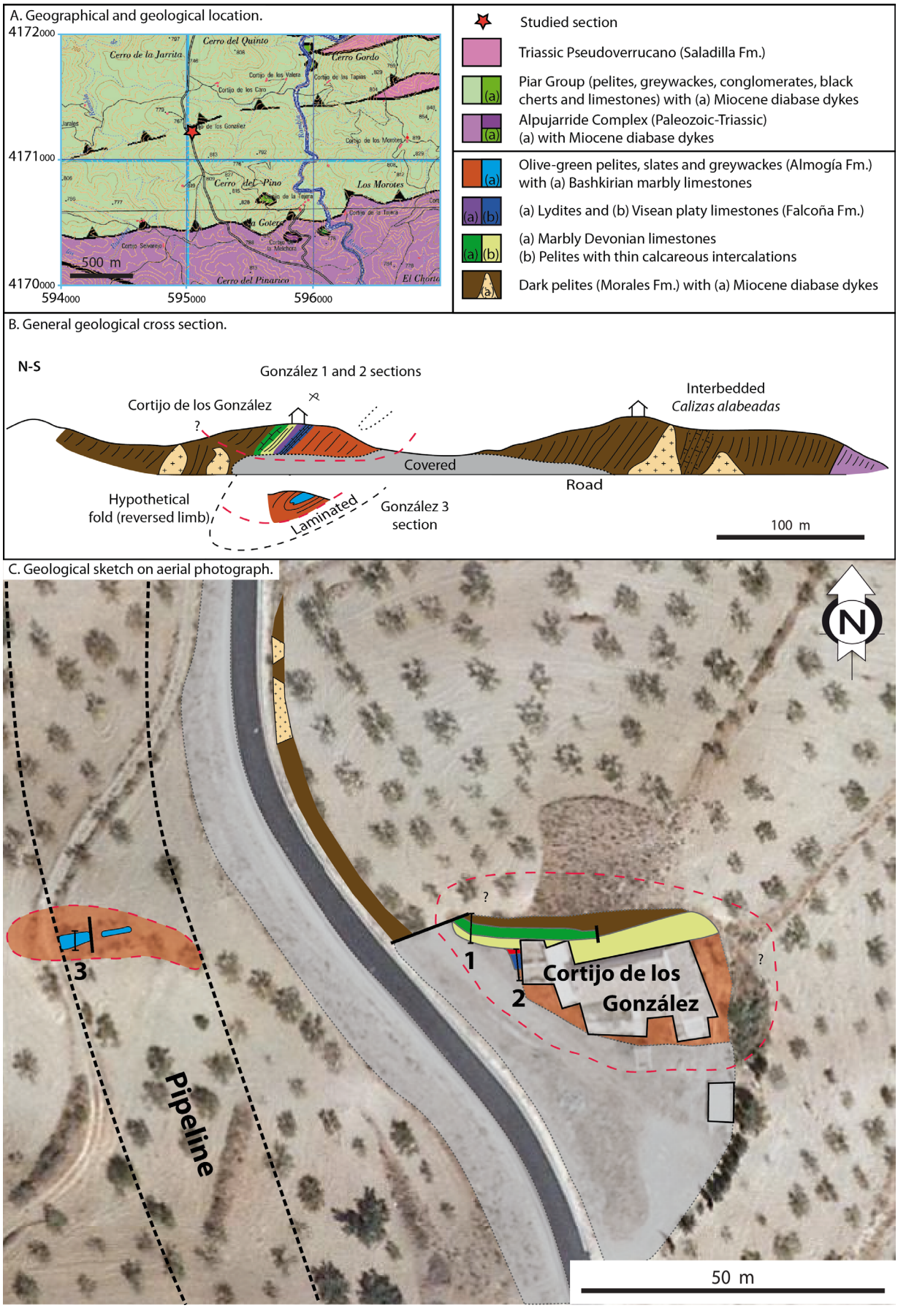


Figure II.1.8. Cortijo de los González outcrop.

brownish pelites and metasandstones crosscut by dykes of diabase rock can be seen along the road a few tens of metres to the north of the farmhouse (Fig. II.1.11.b).

The González 1 and 2 sections crop out along the road from Henares to La Parroquia just close to the western wall of the farmhouse, and four metres apart from each other. The succession is dipping to the N (Fig. II.1.11.a). The González 1 section starts with brown pelites attached to a marbly limestone bed around 2 m thick that yet preserves crinoid bioclasts but that is also strongly foliated (Fig. II.1.11.d). These limestones have released Middle Devonian conodonts (sample 05A-110: location in Fig. II.1.9.A, see paragraph II.1.4. for taxonomic details). Geometrically below, there appear platy limestones with greenish and pinkish pelite intercalations (Fig. II.1.11.c). The marble bed, which outcrops continuously for around 10 m, suddenly disappears towards the east due to a small fault and, to the north of the farmhouse, it is replaced by alternating calcareous slates including cm-thick and strongly folded limestone plates.

After a partially covered shear zone, the González 2 section crops out. Few metres of strongly folded radiolarian chert (lydites) with pelitic partings are visible just below a stonewall of the farmhouse (Fig. II.1.9.A) and, attached to the lydites, there appear very thin-bedded limestones around 1 m thick (Fig. II.1.11.e and f). These limestones yielded Visean conodonts (sample 06P-90: location in Fig. II.1.7.A, see paragraph II.1.4. for taxonomic details), demonstrating that both these limestones and the lydites belong to the Falcoña Fm. The González 2 section ends with green-olive pelites with thin-bedded and fine-grained greywacke intercalations, which probably belong to the Almogía Fm. Consequently, the available litho- and biostratigraphic data indicate that this succession is upside-down, as the conodont associations found are stratigraphically reversed.

The González 3 section was located a few tens of metres to the west of the farmhouse (figs. II.1.9.B, C and II.1.12.a-c). Before the installation of the pipeline (Fig. II.1.10, photo 2005) its western end was formed by a very tight synclinal fold with a subvertical axis (Fig. II.1.12.d and e). Marbly

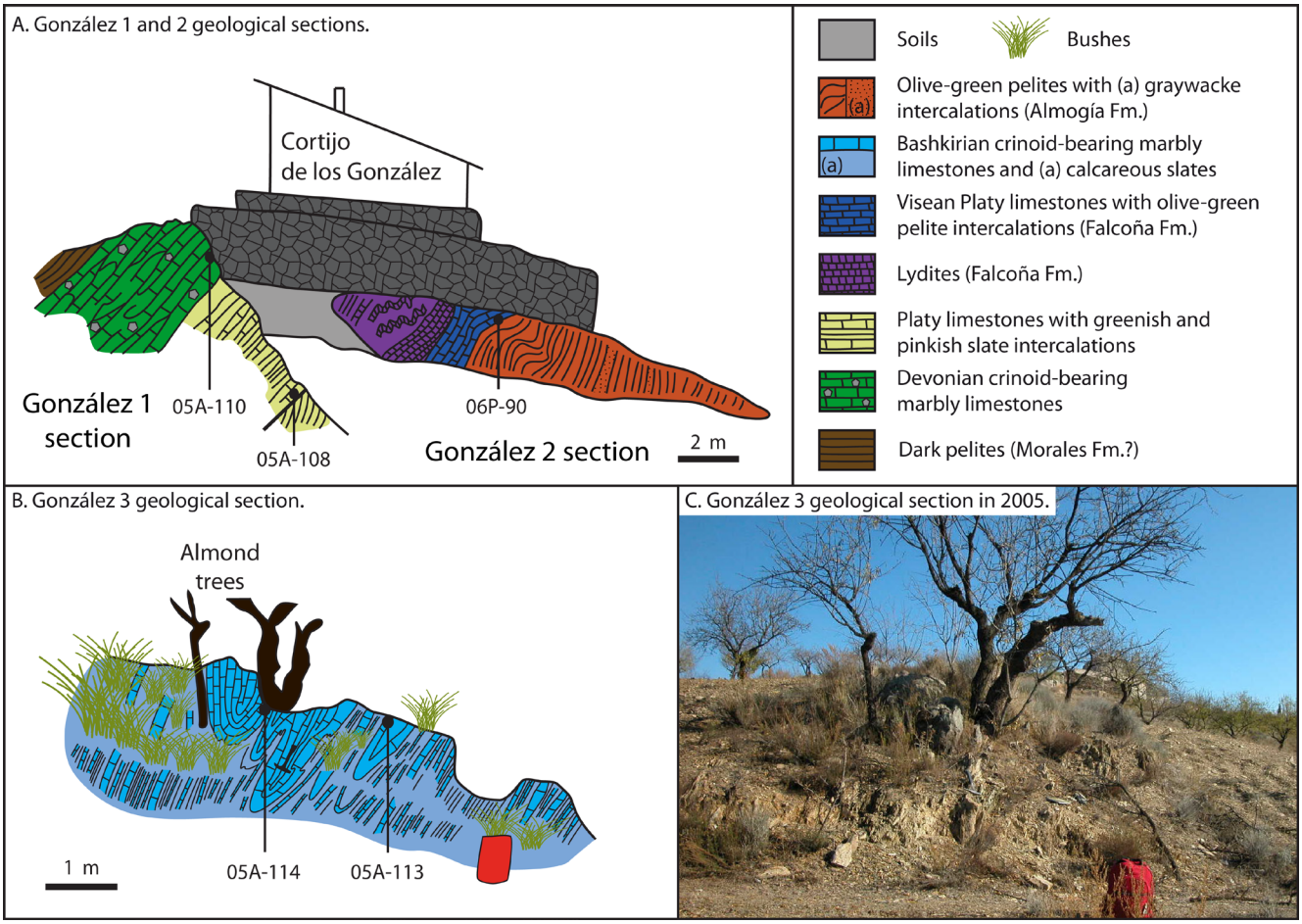


Figure II.1.9. Sections measured and sampled at the Cortijo de los Gonzalez outcrop (see Fig. II.1.8 for location).

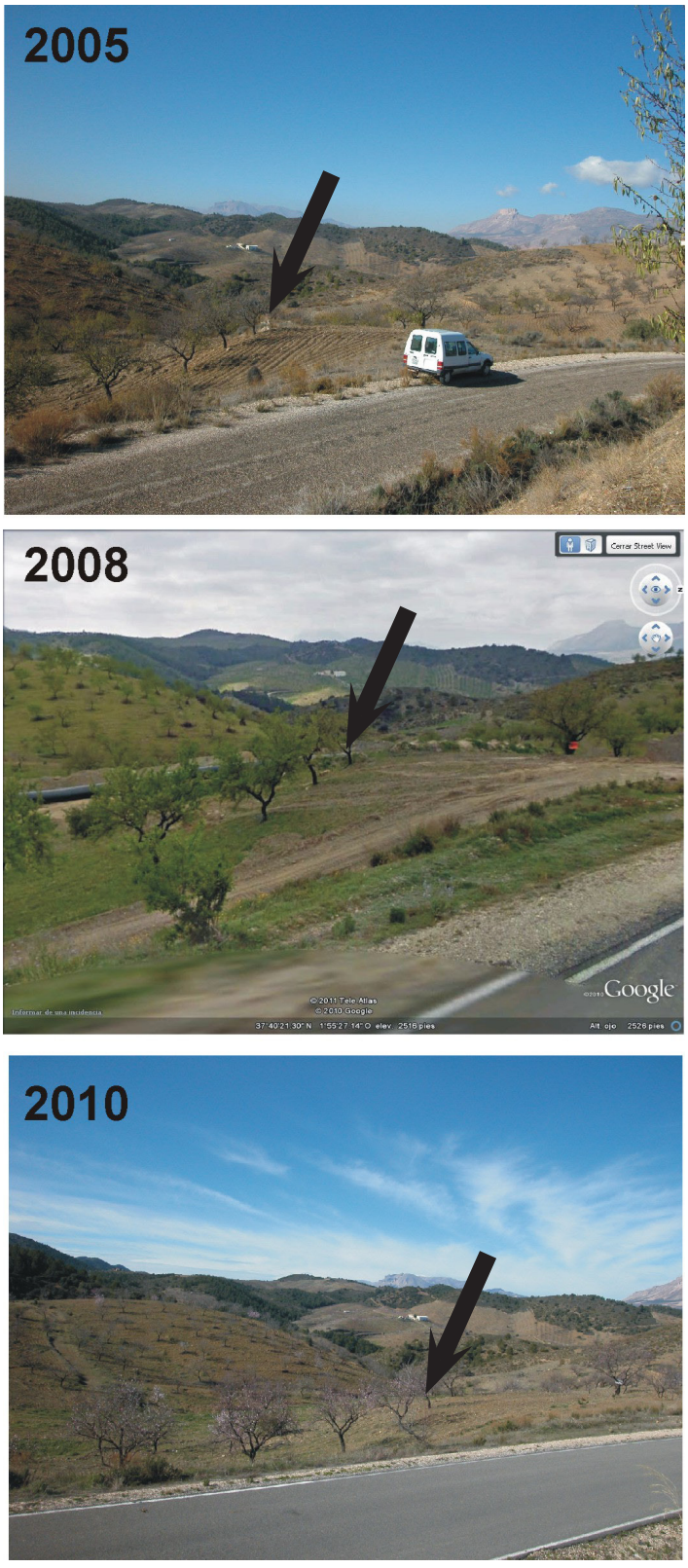


Figure II.1.10. González 3 section before, during and after its destruction.

medium-bedded limestones with crinoid bioclasts (Fig. II.1.12.f) occured in the core, and were surrounded by ploughed calcareous slates and limestones with isoclinal folds. Two limestone samples yielded the first Bashkirian conodonts found in the Betic Cordillera (samples 05A-113 and 05A-114: location in Fig. II.1.9.B). It is yet possible to sample isolated blocks of these limestones in the middle of the cultivated fields, but they are now out of their formerly very complicated, but visible, structural and stratigraphic context. The preliminary data have been published during the course of this dissertation (Navas-Parejo *et al.*, 2012b, see appendix I).

From a structural point of view, the reversed succession of the sections 1-2 is interpreted to be related to the inverted limb of one southward vergent tight synclinal (Fig. II.1.8.B) whose hinge would have been exposed in the González 3 section before its destruction, as in this part of the Cortijo de los González outcrops youngest limestone beds were dated by means of conodonts. After folding, the normal limb of this synclinal was probably refolded, as the hinges in the western end of the core were strongly dipping or nearly vertical, before being tectonically sheared off, a fact that has destroyed the southern, normal limb in the Devonian rocks. As described above, south and north of the Cortijo de los González, the Paleozoic rocks of the Epimetamorphic Malaguide Unit are characterised mainly by pelites and greywackes with diabase dykes including a decimetre-thick intercalation of black fine-grained limestones similar to the *Calizas alabeadas* facies (Fig. II.1.8.B), which are lithologically identical to the Morales and Santi Petri fms., respectivelly, in the Lower Malaguide Unit. On the contrary, the González successions include Middle Devonian marbly limestones which have not been found elsewhere in the Lower Malaguide Unit. Therefore, as it is clearly detached on underlying rocks, and as its stratigraphic succession is strongly folded and is cut off at its base, this Devonian-Carboniferous outcrop is interpreted to be allochthonous on the underlying Paleozoic rocks.

Although Triassic rocks are absent in its neighbourhood, and as its grade of tectonic deformation is very important, the possibility that the González outcrop would have been affected by both pre-Alpine and Alpine tectonics is quite probable.

Figure II.1.11. González 1 (a-d) and González 2 (e-f) sections. ***a****, Partial view of the section (reversed succession). From left to right: Devonian limestones, pelites with limestone intercalations, lydites with pelitic partings and platy limestones (Falcoña Fm.);* ***b****, Slates of the Morales Fm. (left) with diabase dykes (right);* ***c****, Sample location in the recrystallized Devonian limestones (with Eifelian conodonts: sample 05A-110);* ***d****, Detail of the foliation and intrafolial folds in the strongly deformed and recrystallized Devonian limestones (Geotwitty is 3 cm high);* ***e****, Lydites of the Falcoña Fm;* ***f****, Limestones of the Falcoña Fm (with Visean conodonts: sample 06P-90).*

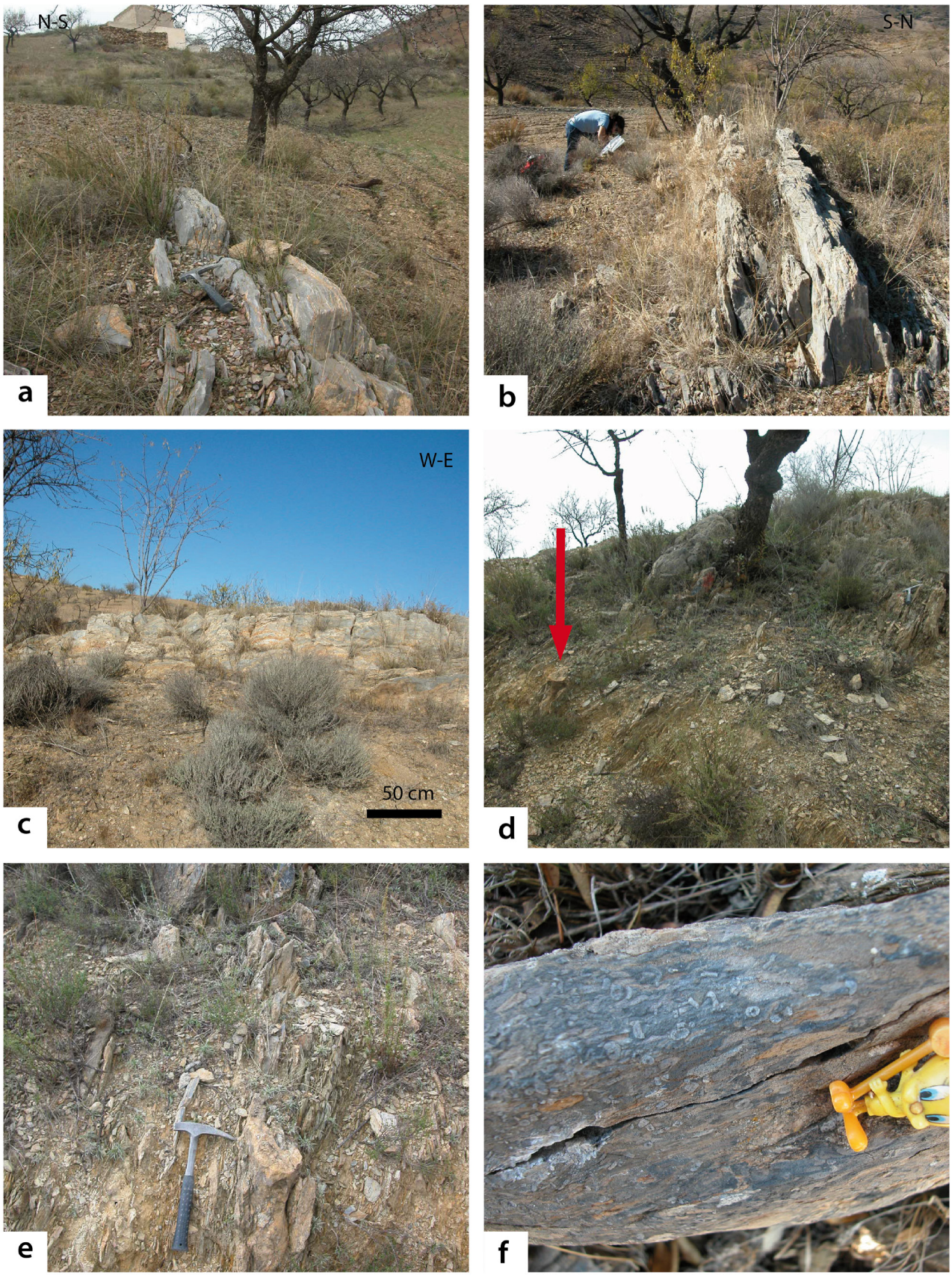


Figure II.1.12. González 3 section (before its destruction in 2008). ***a****, Eastern end of the outcrop partial view, from west to east, with the farmhouse in the background;* ***b****, Eastern end of the outcrop partial view from east to west; the almond tree in the bottom is the same as that viewed in picture d and in Fig. II.1.9.B-C;* ***c****, Partial view of the section from south to north;* ***d****, Western end of the section, with limestone beds strongly folded;* ***e****, Detail (see narrow in d) of the limestones and pelites with subvertical isoclinal folds;* ***f****, Crinoid-bearing limestones, strongly folded, which have yielded Bashkirian conodonts (samples 05A-113 and 05A-114) (Geotwitty is 3 cm high).*

II.1.3.2. Lower Malaguide Unit

As mentioned before, the Cerro Colorado Unit is the most widely outcropping Malaguide unit in the Vélez Rubio-Lorca Corridor, and many field observations have been made in it. The most interesting Paleozoic outcrops found in this area are the following, from east to west: Barranco de Fuente Alegre area, Barranco de los Lobos area, Taberno road section and other outcrops in the area to the south of El Castellón, Barranco Pedrera area, and the region to the SE of Venta Quemada. Some more detailed information on each of these outcrops is provided below.

In the westernmost sector of the Vélez Rubio-Lorca Corridor, the field observations made on the Malaguide Paleozoic have been concentrated in the area to the south of Venta Quemada and Las Vertientes, which entirely corresponds to this Lower Malaguide Unit (Figs. II.1.4). This area can be reached from the A-92N highway, exit 369.

Bluish Alpujarride phyllites occasionally topped by Triassic carbonates widely outcrop at the footwall of the Lower Malaguide Unit, without any evidence of the presence of the Epimetamorphic Malaguide Unit. The lowermost Paleozoic beds in this area are brown pelites with greywacke intercalations that extensively crop out to the south of the Rambla Bermeja, between the Cortijos de la Bermeja and de la Parra. These brown pelites are strongly sheared and recrystallized (slates), and belong to the Morales Fm. Near to the top of this lithostratigraphic unit, on the southern side of the Rambla Bermeja (around 300 m to the south of the Cortijo de la Parra) there appear the only outcrop detected, up to now, in the Vélez Rubio-Lorca Corridor, of the typical lithofacies that constitute the member of the Morales Fm. known as *Conglomerado de cantos estirados* (Conglomerate with stretched clasts: Fig. II.1.13.a). In the same area, the Santi Petri Fm. widely crops out, with its typical *Calizas alabeadas* facies, in the hills located on the northern side of the Rambla Bermeja near to the Cortijos de la Parra and la Pililla. Further northwest, along the Rambla Cañada Hermosa, the Santi Petri Fm. broadly appear for more than 3 km along the river course (Fig. II.1.13.b-d), and, mainly, from the Cortijo de Malagón towards the east, in a belt of several hills partially covered by Quaternary alluvial deposits. No evidence of the Falcoña Fm. has been found, however, probably due to strong tectonics coupled with limited exposure along the watercourse. Nonetheless, good but strongly tectonized outcrops of the Almogía Fm., mainly with pelite-greywacke facies (Fig. II.1.13.e and f), but also locally conglomeratic, have been observed to the E of the Cortijos de la Bermeja, and to the SE of El Matián, where the existence of several (up to now unknown) outcrops of Tertiary rocks probably belonging the Ciudad Granada and Viñuela groups has also been detected.

From the west to the east along the corridor, the Morales Fm. crops out discontinuously at the base of the Lower Malaguide Unit, as it is from time to time (as well as the overlying Santi Petri Fm.) tectonically laminated along the shear zone on top of the underlying Alpujarride Triassic rocks. In addition, from Vélez Rubio towards the east, the rocks of the Morales Fm. are crosscut by a system of dykes of basic subvolcanic rocks, as it is in the classical outcrop area of the Morales Fm. in the Montes de Málaga. Good outcrops of this formation can be found to the south of the El Castellón of Vélez Rubio and, especially, to the east of this town, in the Umbria de Fuente Alegre area, where several sections have been studied and sampled along the Barranco de los Lobos and Barranco Hondo. In particular, along the Barranco de los Lobos (Fig. II.1.14), a section made of more than one hundred of metres of very thin-bedded dark slates with cm-thick to a few m-thick dykes of basic subvolcanic rocks belonging to the Morales Fm. has been studied (Fig. II.1.15.a and b). To the north of the Cortijo del Álamo (A-92N highway, exit 412), these slates intercalate a few m- to cm-thick conglomerate beds (Fig. II.1.15.c and d). These conglomerate beds are usually fine-grained and have a siliciclastic composition. Nonetheless, upwards in the succession, several much coarser-grained and thicker bedded conglomerate beds also include abundant yellowish and greyish to bluish limestone pebbles (figs. II.1.14.C and II.1.15.e). Some of such limestone pebbles have been sampled for conodonts, with no concluding results, unfortunately. In spite of the stratigraphic position, near to the top of the Morales Fm., a few tens of metres below the first beds undoubtedly belonging to the Santi Petri Fm, which is widely outcropping towards the N, these conglomerate beds show a facies totally different from that of the Conglomerates with stretched clasts member that regionally characterise the top of the Morales Fm.

To the NW, the typical *Calizas alabeadas* facies crop out for nearly 2 km along the Barranco de los Lobos riverbed (Fig. II.1.15.f) in normal succession dipping northwards and reaching a thickness

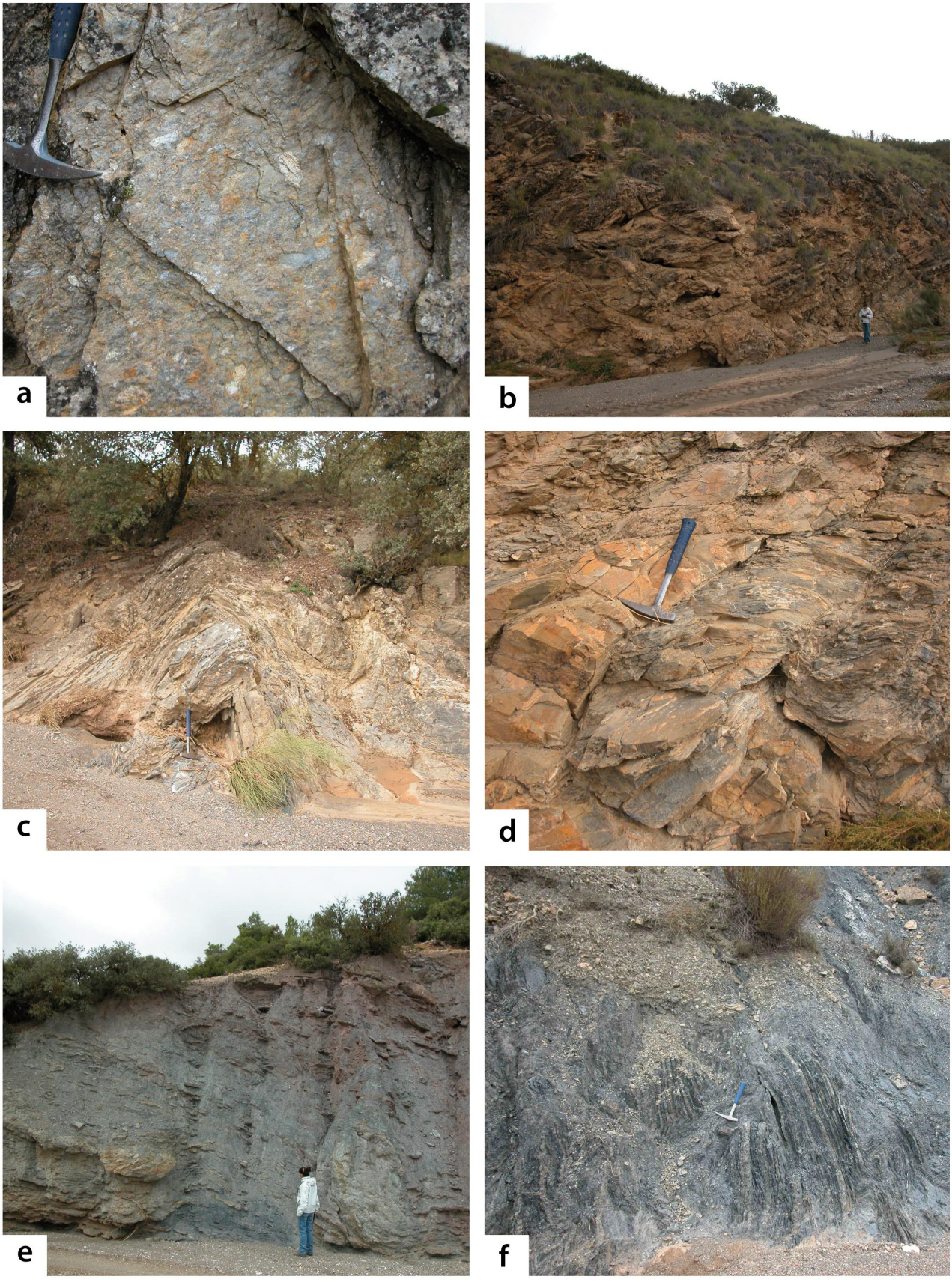


Figure II.1.13. Western Vélez Rubio-Lorca Corridor. ***a****, Conglomerate with stretched clasts (top of the Morales Fm.) in the Rambla Bermeja;* ***b****, Calizas alabeadas (Santi Petri Fm.) in the Rambla Cañada Hermosa;* ***c****, Fold in Calizas alabeadas;* ***d****, Detail of fold in picture b;* ***e****, and* ***f****, Bluish slates of the Almogía Fm. (Culm facies) in the Rambla Bermeja.*

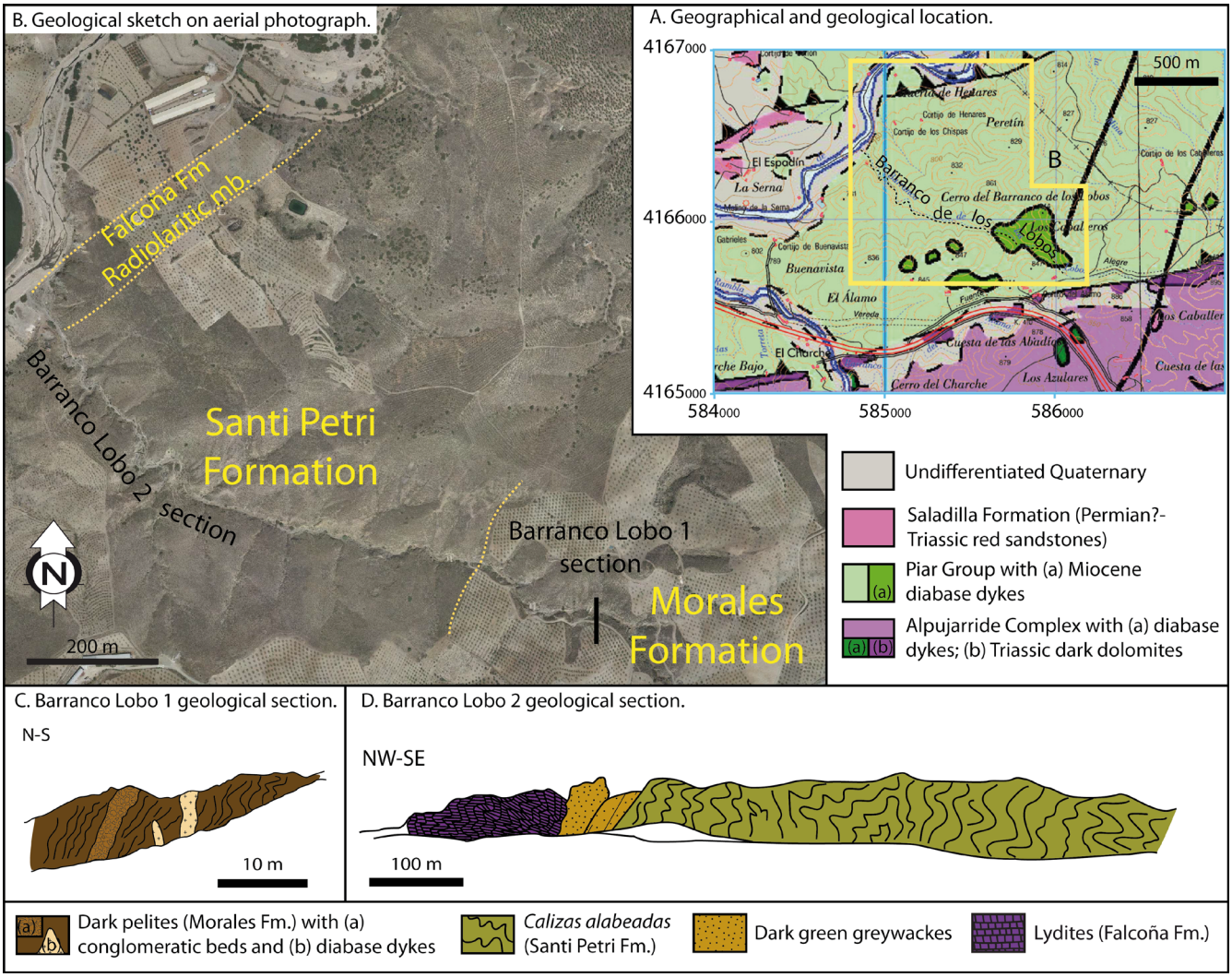


Figure II.1.14. Barranco Lobo outcrop.

of more than 500 m. The calcareous turbidites of the Santi Petri Fm. preserve abundant sole markings and incomplete Bouma-sequences, and are strongly deformed, showing also here typical warps or *alabeos*. Where the barranco reaches the Rambla de Chirivel (from here downwards named Río Cornejos), the Santi Petri Fm. ends with a few tens of metres of dark red- to brown reddish-coloured slates that intercalate thin sandstone beds and strongly deformed thick-bedded greywackes. Close to the confluence with the river, the top of the Barranco de los Lobos succession is made by about 70 m of strongly folded black chert that form a near continuous belt, outcropping for more than 2 km in a ENE to WSW orientation, which was already mapped by Roep (1972). Despite its upper limestone member is totally absent, these cherts certainly belong to the Falcoña Fm, as they are found northeastwards in the Barranco Hondo where it has been well characterised and dated, as it will be described below with more detail.

The same belt of the Lower Malaguide Unit outcropping along the Barranco de los Lobos can be followed, with similar structural and stratigraphic features, toward the east, in the Fuente Alegre area, and toward the west, where good sections of both the Morales and Santi Petri fms. can be observed along the Rambla del Charche, the Taberno road and the Ramblas de la Mata and del Centeno, to the south and SW of El Castellón, respectively. In particular, a good outcrop of the Santi Petri Fm. can be observed along the roadcut of the road AL-7101 from Vélez Rubio to Taberno, about 500 m to the south of El Retamar (Fig. II.1.3 and II.1.7). Here, on top of the bluish phyllites of the Alpujarride Complex, the Paleozoic succession starts with about 50 m of strongly deformed pelites and graywackes of the Morales Fm. Northwards, the *Calizas alabeadas* facies crop out, dipping to the north, for more than 200 m along the road (Fig. II.1.16.a and b). Discontinuous outcrops of the Falcoña Fm., and also of the overlying Almogía Fm. can be laterally followed, close to the Devonian beds sometimes

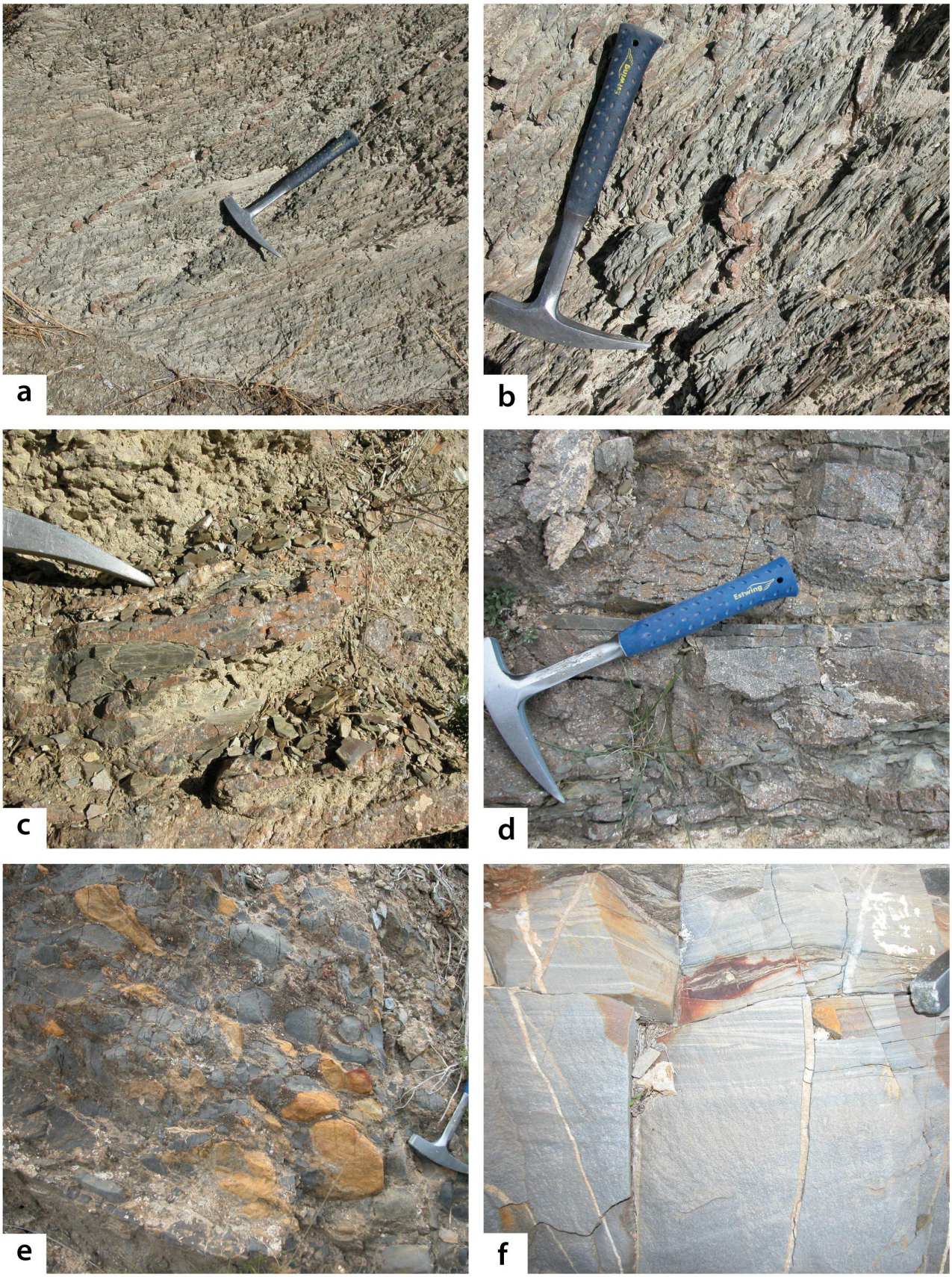


Figure II.1.15. Barranco Lobo outcrop. ***a****, and* ***b****, Slates of the Morales Fm. with cm-thick diabase dykes;* ***c****, and* ***d****, Microconglomeratic beds with erosional base in the Morales Fm.;* ***e****, Detail of a conglomeratic bed with mostly carbonate pebbles;* ***f****, Detail of a turbiditic sequence in the Calizas alabeadas of the Santi Petri Fm., with cross bedding (Bouma sequence interval c).*

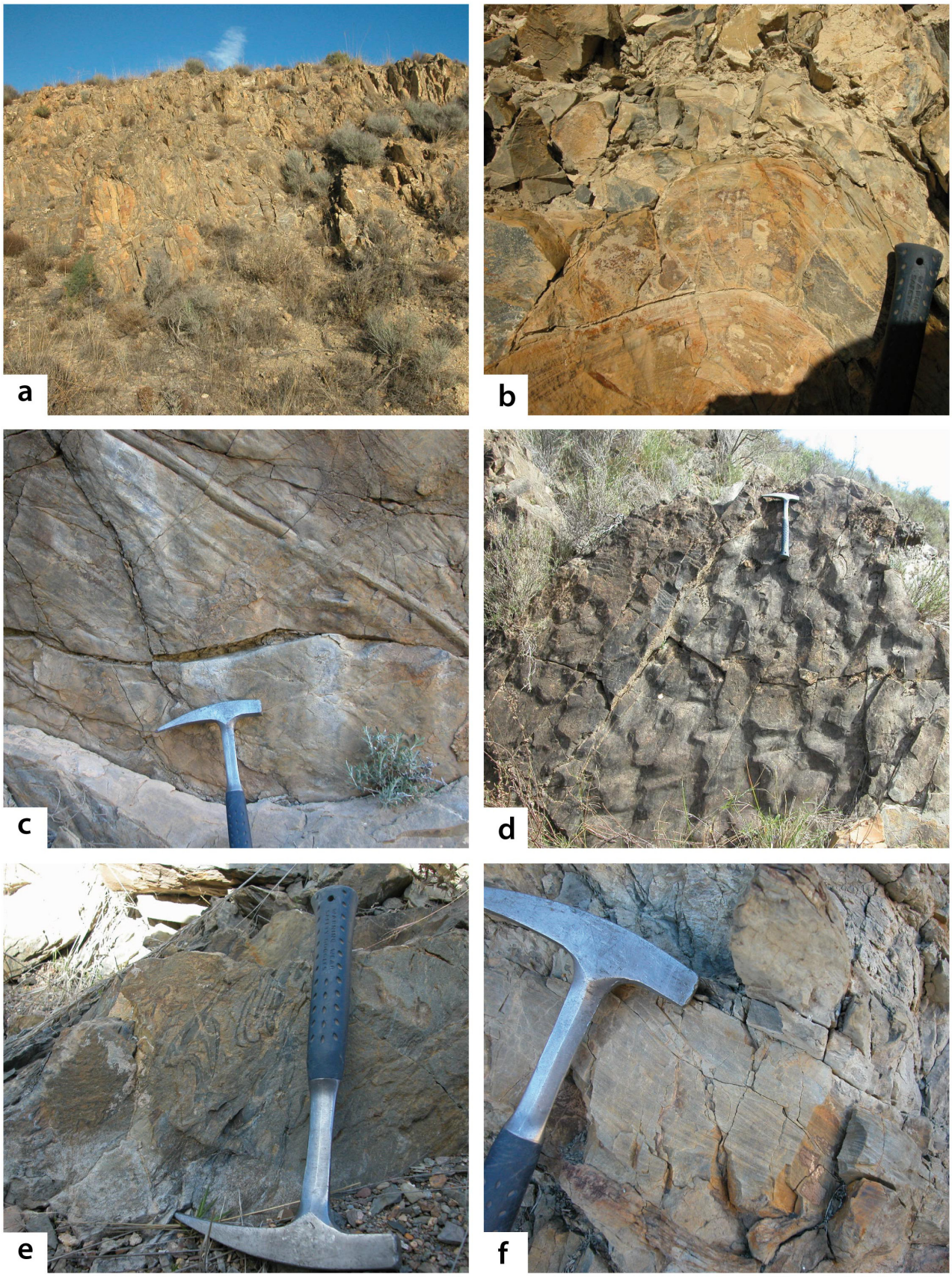


Figure II.1.16. The Santi Petri Fm. in the Taberno road (a-b) and Barranco Fuente Alegre (c-f). ***a****, General view of Calizas alabeadas facies;* ***b****, Detail of a fold in picture a;* ***c****, Sole markings (groove casts) in a turbiditic bed of Calizas alabeadas;* ***d****, Ripples in the top of a Calizas alabeadas turbiditic bed;* ***e****, Microslump in a bed of Calizas alabeadas;* ***f****, Detail of a turbiditic sequence.*

associated with varicoloured pelites and platy limestones bearing dacryoconarids that delineate the contact with the Upper Malaguide Unit to the SW of the El Castellón hill (figs. II.1.3 and II.1.7).

Between Chirivel and Vélez Rubio, the *Calizas alabeadas* frequently disappear between pelitic facies, the underlying one assignable to the Morales Fm., and the overlying one attributable to the Almogía Fm., although to the S of the Cañada de la Saladilla, an important outcrop of Calizas alabeadas also appears (Fig. II.1.7).

Along the Barranco de Fuente Alegre (SE of the Cerro del Piar), the Santi Petri Fm. crops out widely, with many beds preserving spectacular sole marking (groove casts and flute casts), ripples, microslumps and turbiditic sequences (Fig. II.1.16. c-f). To the NW, in the Barranco Hondo (Fig. II.1.17.A and B), the Santi Petri, Falcoña, and Almogía fms. occur, with the Santi Petri Fm. forming the core of a SE-vergent antiformal fold, where three stratigraphic sections have been measured and sampled in its eastern limb (Fig. II.1.17.C). In the western (normal) limb, in the left margin, and overlying the typical *Calizas alabeadas* facies of the Santi Petri Fm., a few tens of metres of medium-bedded and medium to very coarse-grained calcareous sandstones appear (Fig. II.1.18.a). These sandstones show much coarser grain size than turbidites of the Santi Petri Fm. (which are always fine- to very fine-grained) and contain bioclasts of pebble to cobble size, among which one fragment of coral have been recognized in the beds measured in the Fuente Alegre 1 section (Fig. II.1.18.b). Stratigraphically above these northwest-dipping calcareous sandstones, mainly in the river bed of the Barranco Hondo, a few tens of metres of very thin-bedded black radiolarian cherts (lydites) crop out (Fig. II.1.18. c and d). These lydites are strongly deformed and show isoclinal folding. On top of the lydites, on the right margin of the Barranco Hondo and within the southeast

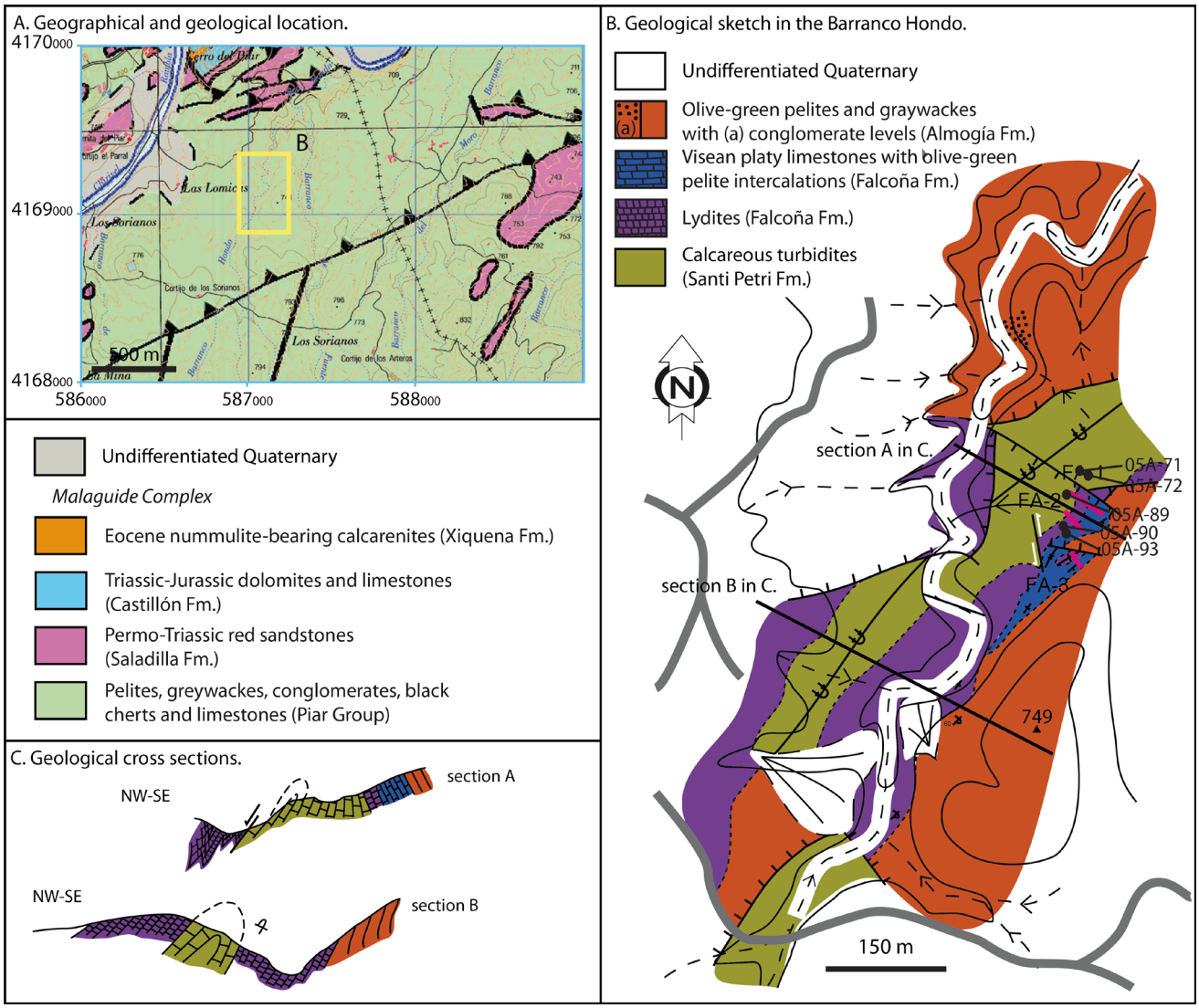


Figure II.1.17. Fuente Alegre west outcrop (Barranco Hondo).

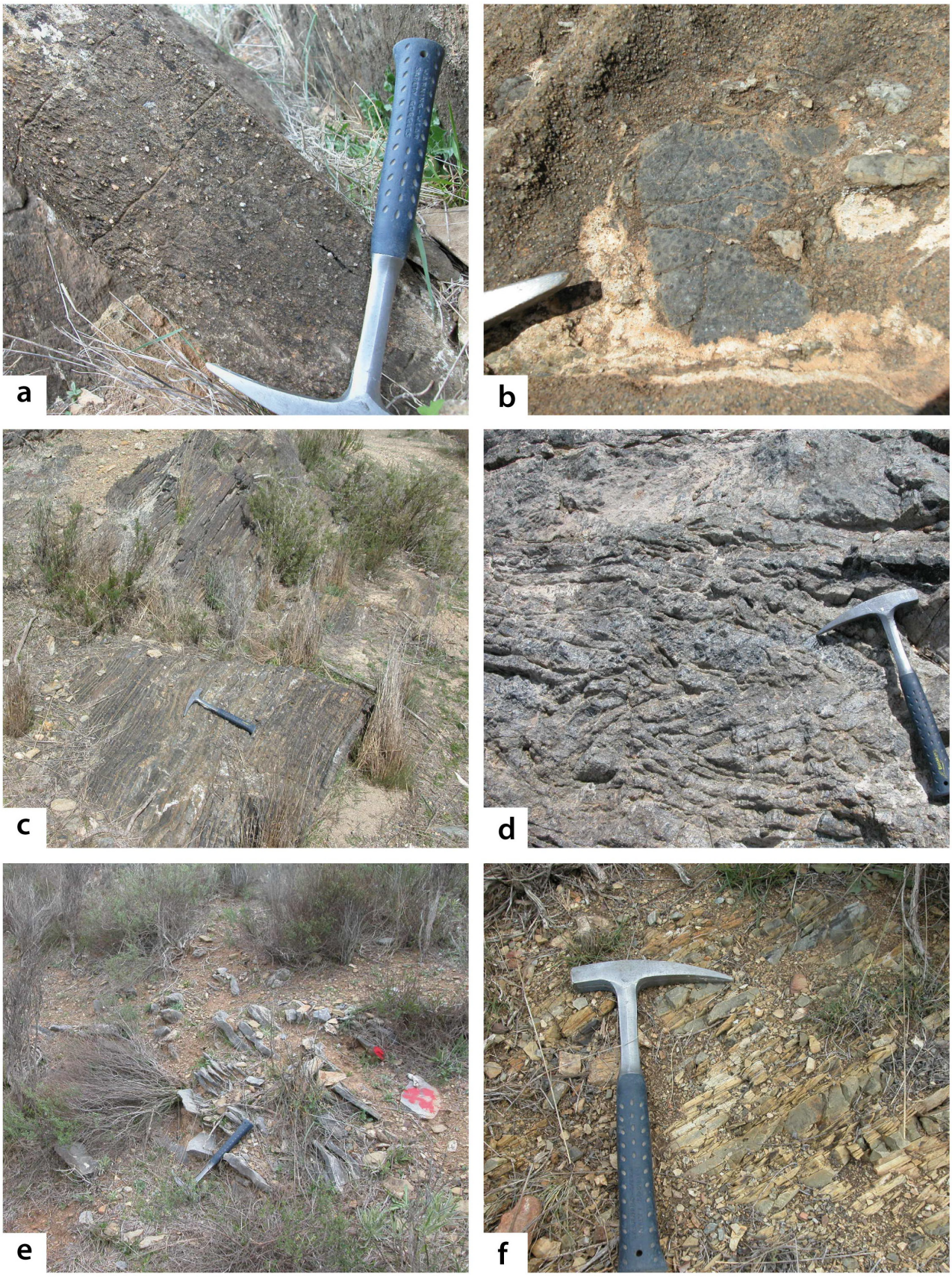


Figure II.1.18. Fuente Alegre outcrop (Barranco Hondo). ***a****, Amalgamation surface in the carbonate turbidites with quartz grains and crinoids that underlie the lydites of the Falcoña Fm.;* ***b****, Clast of coral limestone within the carbonate turbidites in picture a;* ***c****, and* ***d****, Lydites of the Falcoña Fm. (lower member) with isoclinal folds;* ***e****, General view of thin limestone beds of the Falcoña Fm. which have yielded Late Visean conodonts (samples 05A-71, 05A-72, 05A-89, 05A-90 and 05A-93);* ***f****, Platy limestones with intercalated pelites of the Falcoña Fm.*

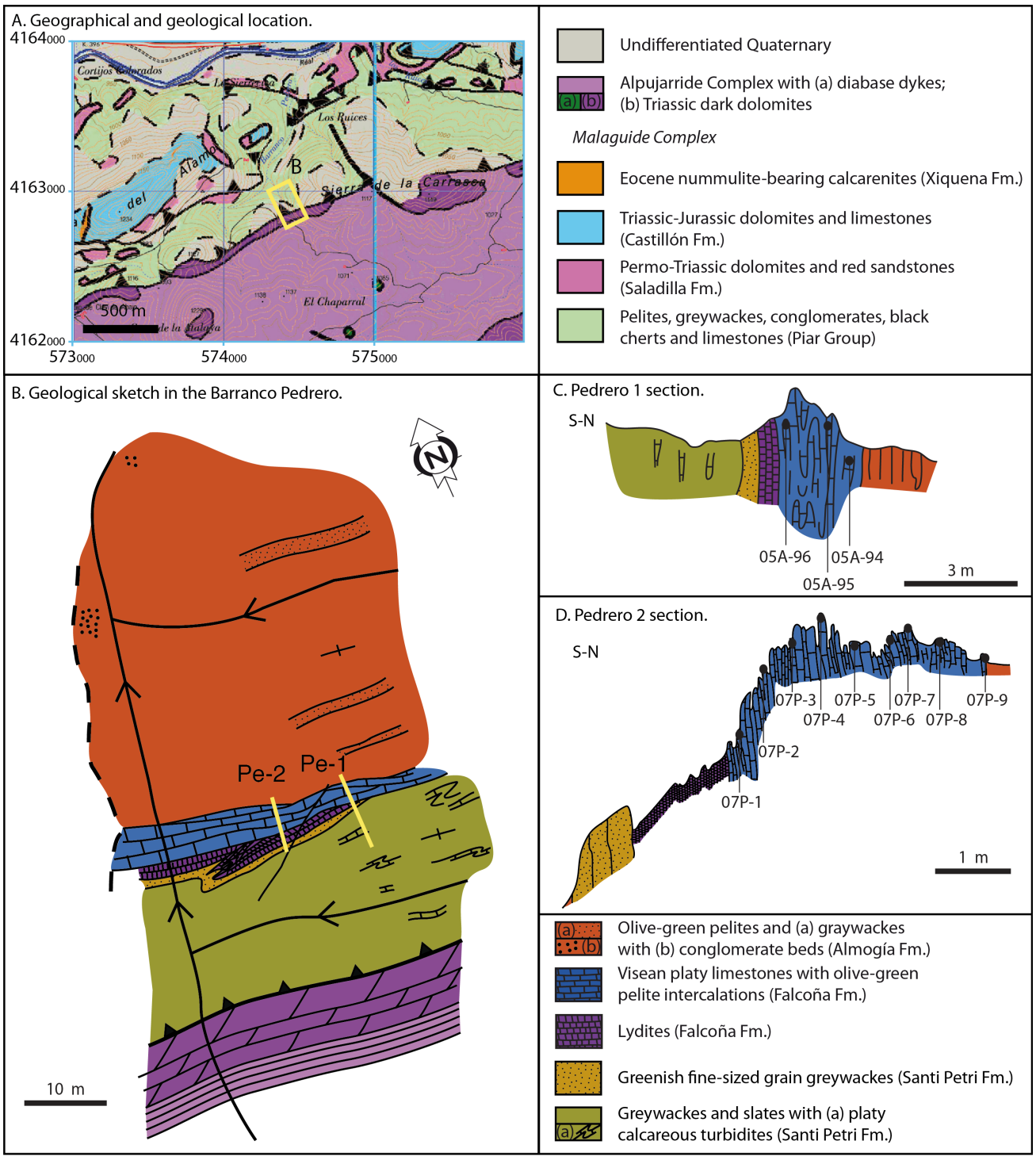


Figure II.1.19. Barranco Pedrero outcrop.

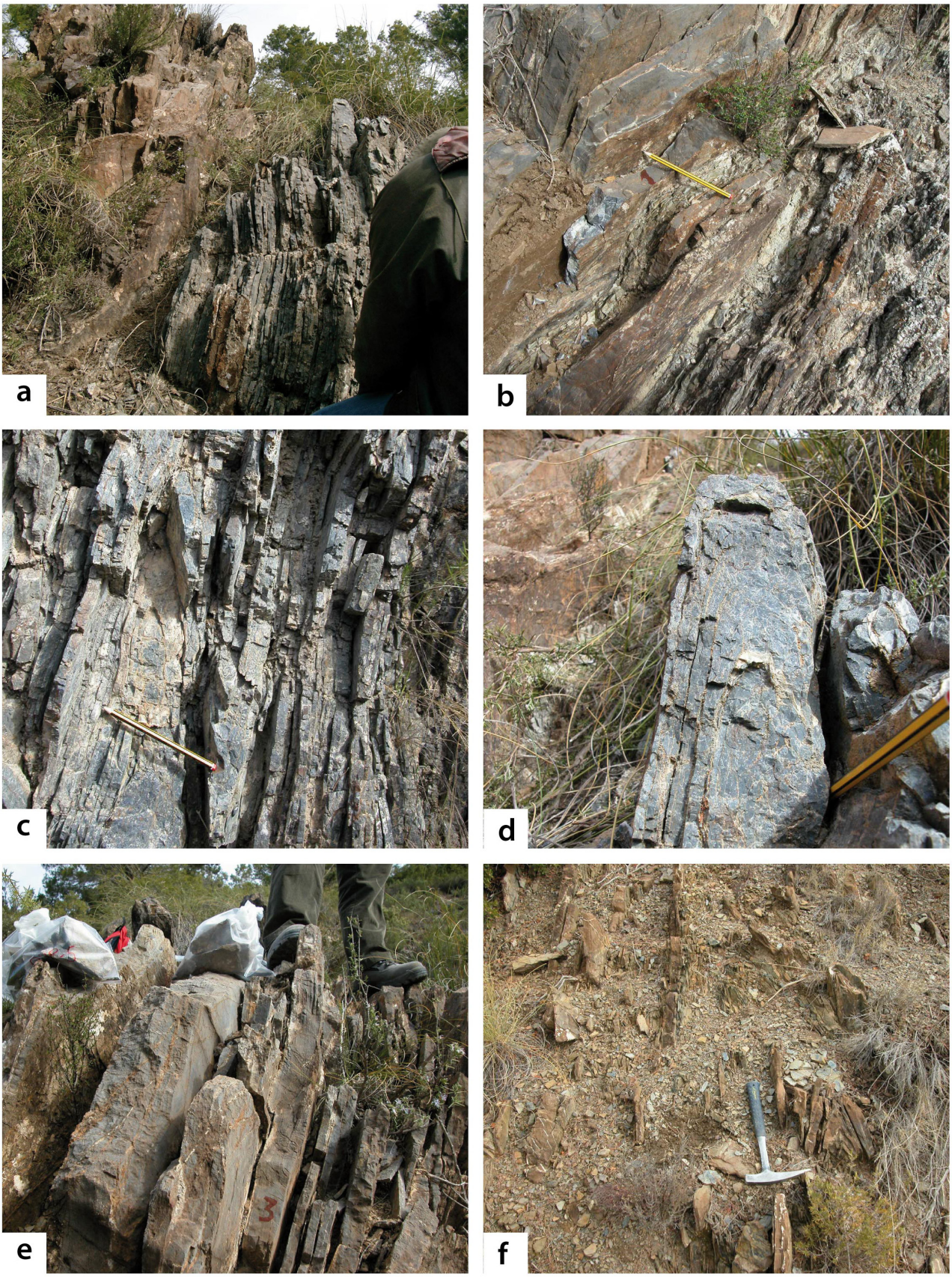


*Figure II.1.20. Pedrero section. **a**, and **b**, Stratigraphic contact between lydites (right) and limestones (left) of the Falcoña Fm.; **c**, Detail of strongly folded lydites of the Falcoña Fm.; **d**, Detail of a isoclinal fold hinge in the lydites; **e**, Platy limestones of the Falcoña Fm. that have released Visean conodonts (samples 05A-94, 05A-95, 05A-96, 07P-1 to 9); **f**, Strongly folded thin bedded greywackes and pelites of the Almogía Fm.*

verging, vertical to locally reversed limb of the fold, a few metres of platy limestones occur on top of the radiolarian cherts and in relation to intercalated dark-coloured pelites (Fig. II.1.18.e and f). These platy limestones were sampled and dated by means of conodonts as Upper Visean (Navas-Parejo *et al.*, 2006, 2008, see appendix I; see also section II.1.4), which confirm that they belong to the Falcoña Fm. This demonstrates for the first time the existence of this formation in the Malaguide outcrops of the Vélez Rubio area, with its two members as defined in the type region near Málaga. Towards the north, and along the Barranco Hondo, dark olive-green pelites and greywackes with conglomeratic beds typical of the Almogía Fm. constitute the highest stratigraphic beds in the normal, northern limb of the antiform that affects the succession of the Lower Malaguide Unit.

The last section studied in this Malaguide Unit is located to the east of the Sierra del Álamo, about 8 kilometres to the SW of Vélez Rubio (figs. II.1.3. and II.1.7). In a tributary watercourse of the Barranco Pedrero, less than 10 metres of very thin-bedded black cherts (lydites) and very thin-bedded greyish-bluish limestones vertically crop out (figs. II.1.19.A and B, II.1.20.a, b and e). The lydites are strongly folded and show the typical isoclinal folding that commonly affects the Falcoña Fm. cherts (Fig. II.1.20.c and d). Moreover, the overlying limestones have yielded Upper Visean conodonts (Navas-Parejo *et al.*, 2008, see appendix I; see also section II.1.4). To the north, stratigraphically over the Visean limestones, dark pelites with very thin greywacke beds of the Almogía Fm. occur, and they continue to the Barranco Pedrero (Fig. II.1.20.f). A few tens of metres to the south of the pelites below the Falcoña Fm., the Paleozoic formations of the Malaguide Complex older that the Falcoña Fm. are tectonically laminated, and they get in direct contact with dark Triassic dolomites of the Alpujarride Complex.

There are many outcrops of the Almogía Fm., including its Retamares member, belonging to the Lower Malaguide Unit, which are located along an almost continue band along the contact with the overlying Upper Malaguide Unit (to the south of the Cañada de la Saladilla, Barranco Pedrero, Cerro de la Monja, to the south of El Castillón and in the Fuente Alegre area). Finally the rocks of the Almogía Fm. are covered by the red pelites, sandstones and conglomerates of the Permian?-Triassic Saladilla Fm., without a clear presence of the Marbella Fm. in any outcrop of the Lower Malaguide Unit.

To summarize, the Paleozoic successions of the Lower Malaguide Unit are mainly characterized by the presence of the Calizas alabeadas of the Santi Petri Fm. showing typical facies, on top of which medium to thick-bedded calcareous sandstones frequently appear. This unit also includes the Morales Fm., crosscut by dykes of basic subvolcanic rocks, the Falcoña Fm., characterized in this unit by the presence of both a radiolaritic and a calcareous members, and the typical coarse- (Retamares member) to fine-grained Culm facies of the Almogía Fm.

II.1.3.3. Upper Malaguide Unit

As mentioned before, in this study I will consider the two uppermost Malaguide Alpine tectonic units of Roep and MacGillavry (1962) as only one Upper Malaguide Unit that is subdivided in many laterally discontinuous tectonic slices. The slices of this higher rank unit were mostly characterised by previous authors by their Meso-Cenozoic deposits. Nonetheless, regarding their Paleozoic successions, significant features also appear in comparison to those of the lower Malaguide units, in particular; 1) the Paleozoic successions are less thick; 2) they lack the Morales and Santi Petri fms; 3) they bear discontinuous limestone horizons of Silurian and Devonian age that have not been found in the Lower Malaguide Unit; and 5) they also present the Marbella Fm., absent in the other two Malaguide units. Field observations have been made mainly in the Vélez Rubio area and the most important ones are described below.

The westernmost section studied in the Upper Malaguide Unit crops out in the Barranco Reolí (Los Gaticos section, see location in Fig. II.1.3 and II.1.7), about 1 km to the NW of the Cerro de la Monja (Fig. II.1.21.A). The section is well exposed in the right bank of the gully, where it starts with a few metres of greenish and reddish pelites forming the core of an anticline (Fig. II.1.21.B). Above them, in the limb

dipping to the north, there appear well-bedded lydites, although the contact is faulted (figs. II.1.21B, II.1.22 a and b). However, in the southern limb of the anticline, the beds are subvertical or slightly dipping south and, overlying the greenish pelites, a few metres of greyish limestones with a cm-thick level of limestone breccia occur (Fig. II.1.22.c and d). Both limestones and breccias were sampled for studying conodonts, but only the breccia level has released an identifiable Upper Devonian fauna (see section II.1.4. for biostratigraphical details). The reddish pelites and the lydites also appear in the left bank of the gully (Fig. II.1.21.B). In brief, although the contact is faulted, the Gaticos succession ends with a few tens of metres of strongly-folded, very thin-bedded black chert, probably belonging to the Falcoña Fm. as the geometrically underlying breccia has yielded Upper Devonian conodonts. Nevertheless, I was not able to find the upper calcareous member of the Falcoña Fm. neither in the left or in the right bank, nor in other gullies in surrounding areas.

To the SW of the Triassic and Jurassic limestones of the El Castellón hill (south of Vélez Rubio), the most complete Paleozoic succession found in the Upper Malaguide Unit crop out. This succession starts with a few tens of metres of platy slate limestones bearing dacryoconarids (Fig. II.1.23.c) followed by a few metres of very thin-bedded black cherts. On top of these lydites, bluish fine-grained and thin-bedded limestones with olive-green pelitic levels have been locally found (Fig. II.1.23.d). They were sampled for conodonts, with no results. In spite of this, I will tentatively assign both the lydites and the bluish limestones to the Falcoña Fm., although Geel (1973) mapped the outcrop in Fig. II.1.23.d as part of her *Tentaculites limestone member*. The Paleozoic succession in this area ends with the typical conglomeratic facies of the Marbella Fm. This conglomerate is composed by shelf limestone (with coral, algae and foraminifer bioclasts), granite, lydite and graywacke pebbles and cobbles (Fig. II.1.23.e).

In the area between Vélez Rubio and the Ermita del Piar to the south od La Tejera hill (Fig. II.1.24.A) an interesting but tectonically complicated Paleozoic succession of the Upper Malaguide Unit can be studied along several gullies. A synthetic geological section is shown in Fig. II.1.24.B. In this area, van den Boogaard (1965) found Devonian conodonts in a limestone block sampled from a coarse-grained conglomerate outcrop, quite probably

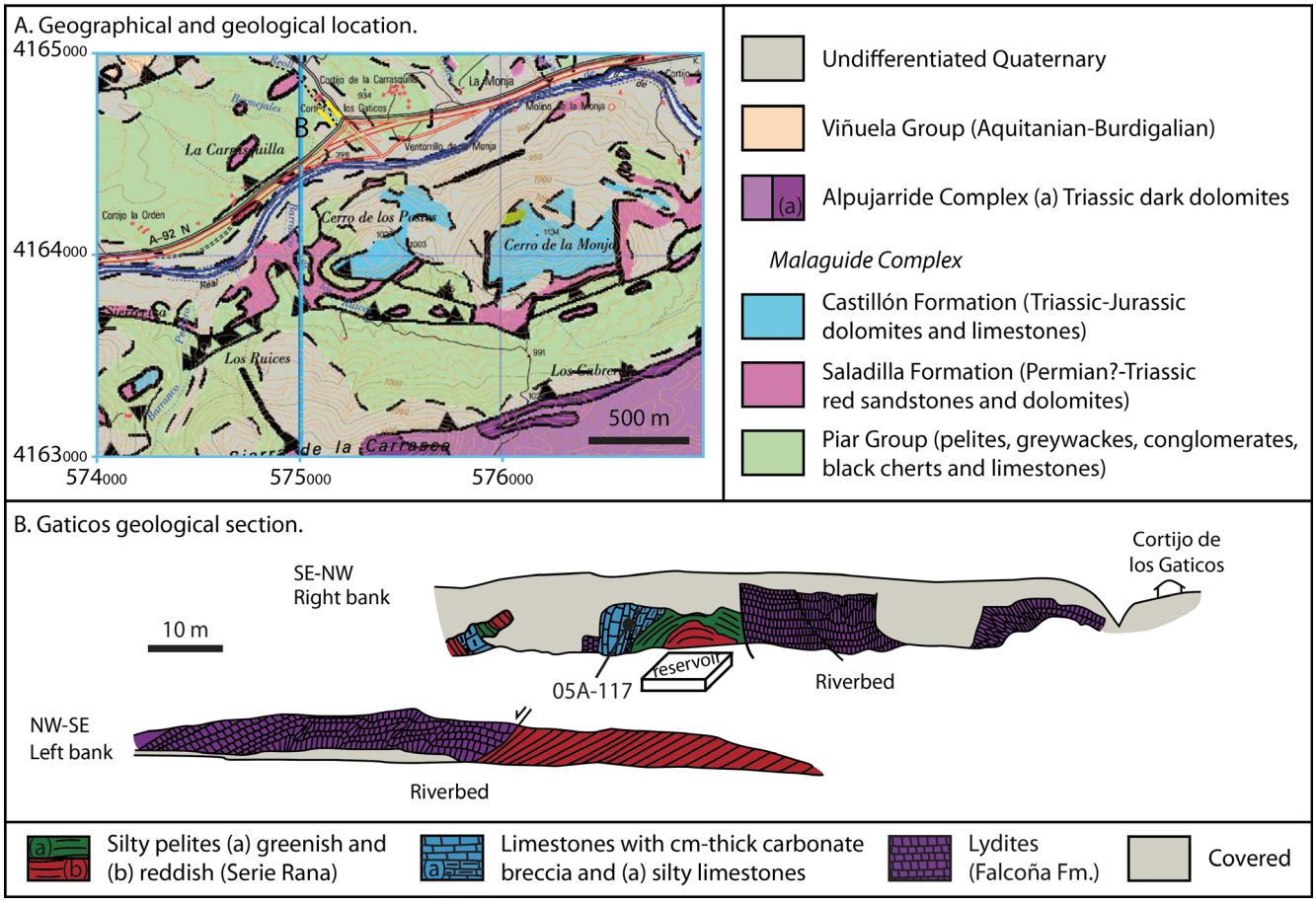


Figure II.1.21. Gaticos section.

Figure II.1.22. Gaticos section (a-d) and La Tejera section (e-f). ***a****, Partial view of the section of the right bank of the Barranco Reolí. From right to left: lydites of the Falcoña Fm., overlying greenish and reddish pelites and limestones although the contact is faulted;* ***b****, Upstream view of the pelitic beds with limestones;* ***c****, Limestones underlying lydites, with a carbonate breccia level;* ***d****, Detail of the carbonate breccia, which had yielded Upper Devonian conodonts (05A-117);* ***e****, Retamares conglomerate of the Almogía Fm. with a boulder-sized imestone clast below the hammer;* ***f****, Kink-folds in slates (Almogía Fm.?).*

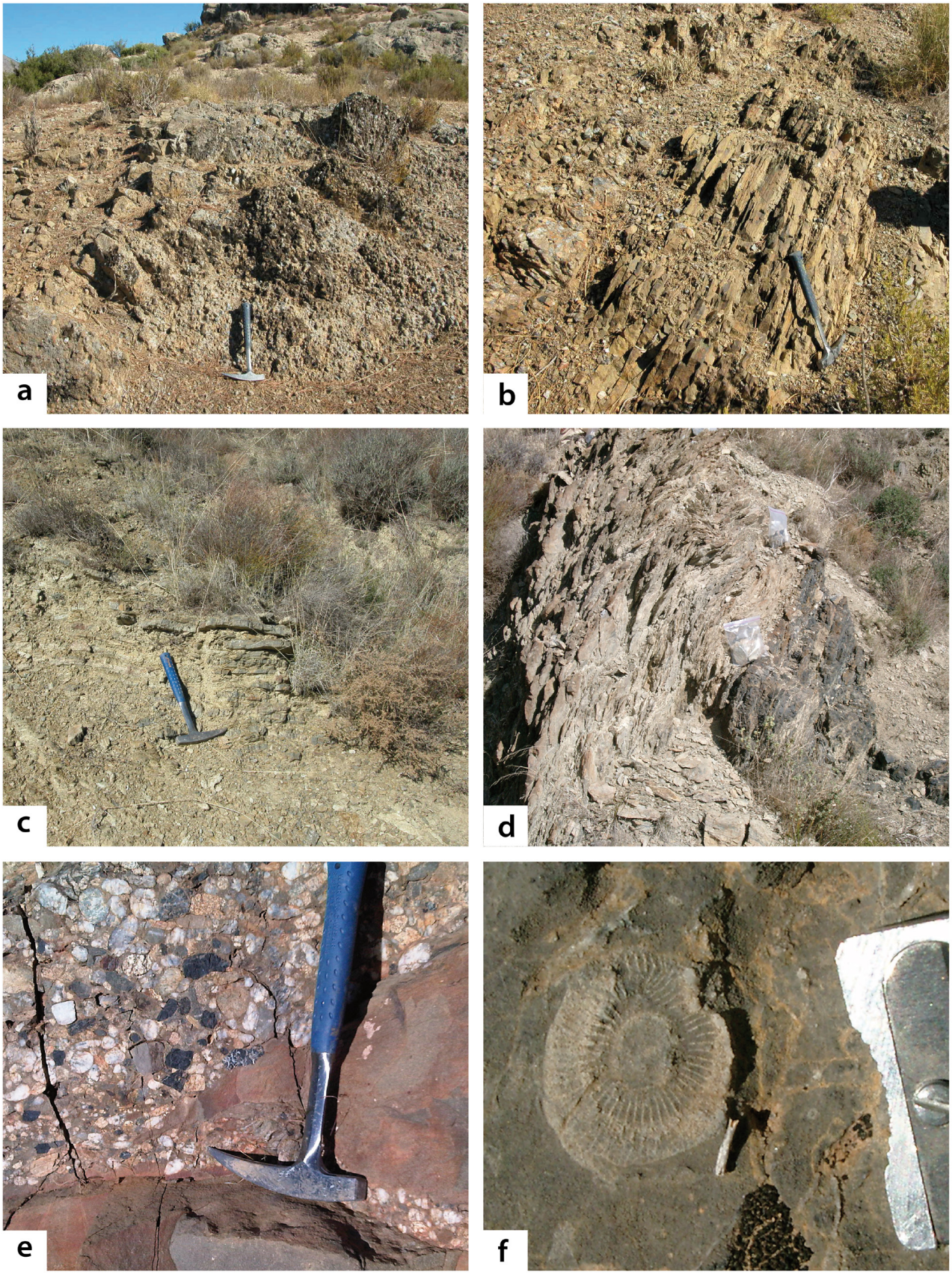


*Figure II.1.23. El Piar section (a-b), SW of El Castellón (c-e) and Rambla de los Coroneles (f). **a**, Retamares conglomerate of the Almogía Fm. (one limestone clast sample here has released Upper Devonian conodonts, sample 06P-87); **b**, Slates of the Almogía Fm. with an overlying Retamares conglomerate bed; **c**, Calcareous platy limestones with dacryoconarids; **d**, Thin-bedded limestones interbedded with pelites on top of black lydites of the Falcoña Fm.; **e**, Conglomerate of the Retamares member in the Rambla de los Coroneles; **f**, Coral bioclasts in a big limestone boulder of the Marbella Fm.*

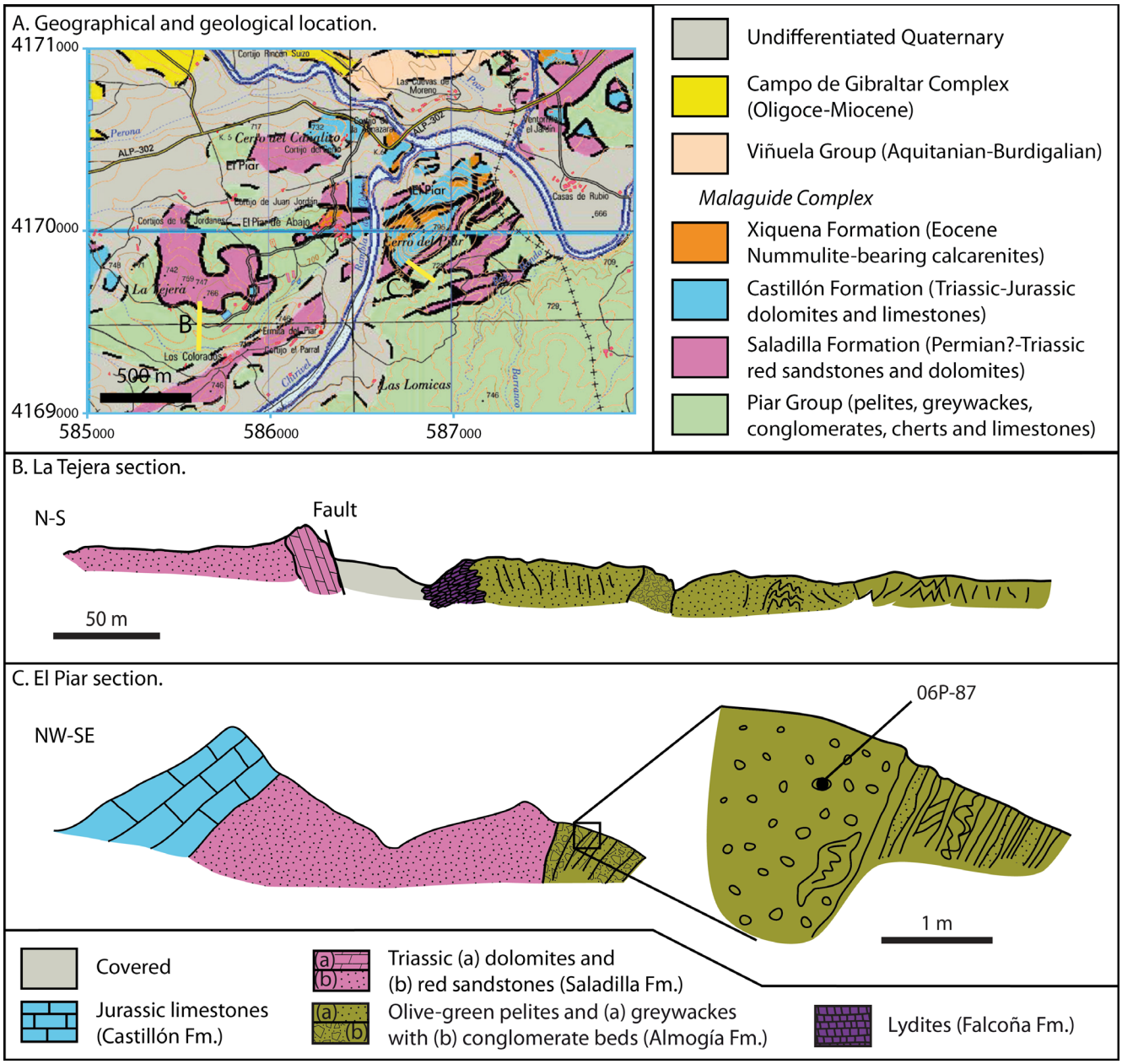


Figure II.1.24. Geologic and geographic location of La Tejera and El Piar sections.

corresponding to that shown in Fig. II.1.22.e, which show equivalent facies to the Retamares conglomerate of the Almogía Fm. This outcrop, from which one limestone block were sampled for conodont analysis with no results, makes part of a strongly deformed subvertical succession, where two parts can be distinguished. To the north of the conglomerate, a badly exposed subvertical tectonic contact bounds the Triassic dolostones and red sandstones and pelites of La Tejera hill from the Paleozoic succession outcropping to the south of it. The latter starts with a few metres of strongly folded and very thin-bedded black cherts, identical to the lydites of the lower member of the Falcoña Fm. Just to the south of the lydites, a subvertical succession essentially made of greywackes with subordinate microconglomerates and olive-green pelites certainly corresponds to the Almogía Fm. To the south and SE of the conglomerate above mentioned, although with unclear geometric relationships (probably tectonic), good outcrops of strongly folded shales (Fig. II.1.22.f) appear for a few hundred of metres along the gullies, probably constitute the core of an antiform. They could correspond to the upper pelitic member of the Almogía Fm., although a Devonian age can not be totally excluded, as they locally show a slightly calcareous character. Actually, these pelites remember those that alternate with thin limestone beds bearing tentaculites in other outcrops of the Upper Malaguide Unit, in particular to the west of Vélez Rubio (see also Fig. II.1.7). Therefore, although a part of the *Greywacke-shale member* of the *Piar Formation* (*sensu* Geel, 1973) could correspond

to the Almogía Fm., another large part of it could belongs to another, yet unnamed, essentially pelitic Devonian formation including thin limestone horizons woth (Silurian? and) Devonian microfossils.

To the south of the Cerro del Piar, a narrow stripe of Paleozoic rocks crop out between Triassic deposits belonging to the Lower Malaguide Unit (to the south) and the Triassic and younger deposits of this Upper Malaguide Unit (Fig. II.1.24.A). These Paleozoic rocks correspond to typical slates and conglomerates of the Almogía Fm. (Fig. II.1.23.a and b). Limestone pebbles from this conglomeratic level (Retamares member of the Almogía Fm.) have yielded Upper Devonian conodonts.

Another interesting outcrop of the Almogía Fm. in the Upper Malaguide Unit was found close to the road from Lorca to La Fuensanta (MU-701). In a tributary of the Rambla de los Coroneles (500 m to the WSW of the Venta del Coronel) about 500 m of Paleozoic rocks crop out between Quaternary deposits and the Neogene Lorca Basin (to the NE). A polymict conglomeratic bed with erosive channelled base appears intercalated between brown-reddish (wheathered) sandstones with greywacke composition (Fig. II.1.23.e). This conglomerate is made up of well-rounded pebbles and cobbles of quartzites, gneisses, granites, lydites, crinoid-bearing limestones and very fine-grained bluish limestones. This conglomerate corresponds to the Retamares member of the Almogía Fm.

In short, the lower part of the Paleozoic succession of the Upper Malaguide Unit is mainly characterised by the presence of a pelitic succession bearing thin limestone horizons. Unlike the *Calizas alabeadas* of the Santi Petri Fm., which seem to be azoic and located in an equivalent stratigraphic position. These thin limestones provide (Silurian? and) Devonian conodonts and other fossils that allow biostratigraphic dating. Regarding the Falcoña Fm., I found the upper calcareous member only in one outcrop (SW of El Castellón) but without biostratigraphic results.

Finally, all the outcrops of the Marbella Fm. in the Vélez Rubio-Lorca Corridor appear within the Upper Malaguide Unit, closely related to outcrops of the Triassic Saladilla Fm. in most localities (Fig. II.1.7).

II.1.4. Conodont biostratigraphy

Seven of the outcrops described above were sampled for conodonts: González (1, 2 and 3), Pedrero, Barranco Lobo, Fuente Alegre, Los Gaticos, La Tejera and El Piar, with a total of fifty samples and 33,6 kg dissolved (see Table II.1). Twenty three of these samples (from the González 1, 2 and 3, Pedrero, Fuente Alegre, Gaticos and Piar sections) have released identifiable conodonts, as detailed below (see figs. II.1.2, II.1.3 and II.1.7, for location of each section). Devonian conodonts have been found in the Epimetamorphic Malaguide Unit (Gonzalez 1 section) and in the Upper Malaguide Unit (El Piar and Los Gaticos sections). Carboniferous conodonts have been obtained from the Epimetamorphic Malaguide Unit (Gonzalez 2 and 3 sections) and the Lower Malaguide Unit (Pedrero and Fuente Alegre sections).

II.1.4.1. Epimetamorphic Malaguide Unit

As mentioned before, in the surrounding area of the Cortijo de los González, three sections have been sampled and studied in detail (figs. II.1.8 and II.1.9; see table II.1 for sample productivity details).

In the González 1 section five samples have been taken but only one has yielded identifiable conodont elements (05A-110). This sample released *Polygnathus* cf. *costatus* Klapper, 1971, which occur in the *partitus-costatus* zones (Eifelian; Middle Devonian). Several *Icriodus* Pa elements were also found but they could not be classify because of bad preservation.

Polygnathus costatus is characterised by strong tranverse ridges on both lateral margins (Fig. II.1.25.1), which are separated from the carina by adcarinal grooves (Ziegler, 1974).

In the González 2 section one sample (06P-

90), from a total of four, has released conodont elements. Among the fragments found in sample 06P-90, a complete but very badly preserved *Gnathodus bilineatus bilineatus* (Roundy, 1926) Pa element has been found (Fig. II.1.25.2). Despite of its bad preservation, part of an inner parapet and an outer platform with aligned nodules attached to the carina can be observed (Perret, 1993). This species ranges from the *bilineatus* to the *bollandensis* zones (late Visean-early Serpukhovian age; Lower Carboniferous).

In the González 3 section, two of three samples (05A-113 and 05A-114) have yielded conodont elements. Sample 05A-113 has released *Idiognathoides sinuatus* Harris and Hollingsworth, 1933. The presence of this species indicates a Bashkirian-lower Moscovian age (Upper Carboniferous).

Idiognathoides sinuatus is characterised by left-sided Pa elements, which show narrow, long platforms with a medial groove that reaches the posterior end, and a ridged inner parapet (Nemyrovska, 1999) (Fig. II.1.25.5 and 6).

Sample 05A-114 has released *Idiognathoides corrugatus* (Harris and Hollingsworth, 1933), *Idiognathoides sinuatus* Harris and Hollingsworth, 1933, *Declinognathodus noduliferus* (Ellison and Graves, 1941) and *Lochriea* aff. *commutata* (Branson and Mehl, 1941). This conodont association ranges from the *sinuatus-sulcatus* to the *sinuosus* zones (late Bashkirian; Upper Carboniferous).

According to van den Boogaard and Bless (1985)

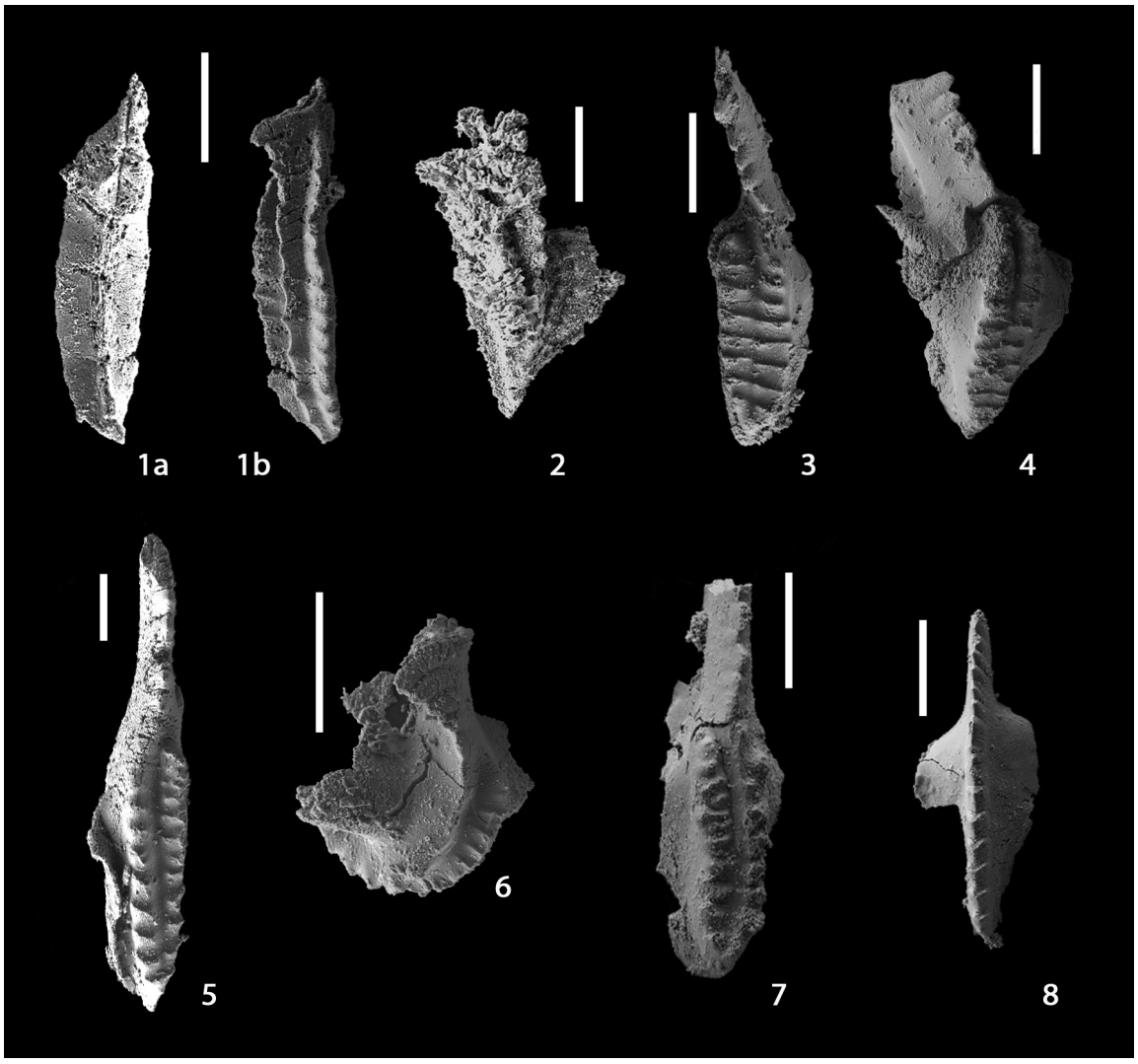


Figure II.1.25. Conodont Plate of González 1 (photos 1a and 1b), González 2 (photo 2) and González 3 (photos 3-8). All are Pa elements. Scale bars are 200 µm.

***1**, Polygnathus* cf. *costatus*; *05A-110-2. 1a, lower view; 1b, upper view. **2**, Gnathodus bilineatus bilineatus; 06P-90-1. **3**, Idiognathoides corrugatus; 05A-114-2. **4**, Idiognathoides corrugatus; 05A-114-5. **5**, Idiognathoides sinuatus; 05A-113-1. **6**, Idiognathoides sinuatus (folded); 05A-114-3. **7**, Declinognathodus noduliferus; 05A-114-4. **8**, Lochriea aff. commutata; 05A-114-7.*

and Nemyrovska (1999), *Idiognathoides corrugatus* is a morphotaxon that corresponds to dextral Pa elements of *Idiognathoides sinuatus*, although they are an asymmetrical pair. *Idiognathoides corrugatus* presents a platform ornamented by continuous parallel transverse ridges without medial groove, and a blade that reaches the outer side of the platform (Fig. II.1.25.3 and 4).

Declinognathodus noduliferous Pa elements are characterised by a blade that reaches the platform centrally, by a short carina declining to the outer parapet, and by a shallow medial groove which gets wider and deeper posteriorly; both outer and inner parapets are parallel and ornamented by nodules (Nemyrovska *et al.*, 2011) (Fig. II.1.25.7).

Specimens here assigned to *Lochriea* aff. *commutata* are carminiscaphate elements with smooth slightly asymmetrical platforms, which are significantly more elongated than in *Lochriea commutata* (Fig. II.1.25.8). They probably correspond to transitional forms that finally evolved to *Declinognathodus orphanus* (van den Boogaard and Bless, 1985; von Bitter and Merrill, 1990).

II.1.4.2. Lower Malaguide Unit

Two different, but very close, sections have been made in the Barranco Pedrero outcrop (Fig. II.1.19). Both sets of samples have released Lower Carboniferous conodonts (see table II.1.1 for sample productivity details).

In the Pedrero 1 section three of four samples yielded conodonts (05A-94, 05A-95 and 05A-96). Samples 05A-94 and 05A-96 have released *Gnathodus joseramoni* Sanz-López, Blanco-Ferrera and García López, 2004. Samples 05A-94 and 05A-95 have also released *Gnathodus bilineatus bilineatus* (Roundy, 1926) and *Lochriea commutata* (Branson and Mehl, 1941). These species coexisted in a short interval in the lowermost *bilineatus* Zone (Visean; Lower Carboniferous).

In the Pedrero 2 section nine of eleven samples have yielded conodont elements. Samples 07P-1, 07P-3, 07P-4, 07P-5, 07P-8, and 07P-9 have released *Gnathodus bilineatus bilineatus* (Roundy, 1926), and samples 07P-2 and 07P-5 also yielded *Lochriea commutata* (Branson and Mehl, 1941). Both species belong to the upper Visean-lower Serpukhovian (Lower Carboniferous). However, the presence of *Gnathodus praebilineatus* (Belka, 1985) in samples 07P-2, 07P-4, and 07P-9, and the presence of *Gnathodus joseramoni* Sanz-López, Blanco-Ferrera and García López, 2004 in samples 07P-3 and 07P-4 allows us to be more specific about the age of these samples, and they belong to the lowermost part of the *bilineatus* zone (Visean; Lower Carboniferous), as in the Pedrero 1 section.

In the Fuente Alegre outcrop, three sections, very close among them, have been studied (Fig. II.1.17). In the Fuente Alegre 1 (FA-1) section, two samples were taken (05A-71 and 05A-72). Both have released *Gnathodus joseramoni* Sanz-López, Blanco-Ferrera and García López, 2004 and *Lochriea commutata* (Branson and Mehl, 1941). As in the Pedrero 1 and 2 sections, the presence of these species indicates the lowermost part of the *bilineatus* Zone (Visean; Lower Carboniferous).

One of a total of five samples (05A-89) has released conodonts in the Fuente Alegre 2 section (FA-2). Sample 05A-89 yielded *Gnathodus bilineatus bilineatus* (Roundy, 1926), *Gnathodus joseramoni* Sanz-López, Blanco-Ferrera and García López, 2004 *Lochriea commutata* (Branson and Mehl, 1941). This conodont association indicates the same age than that in the other sections in this Malaguide Alpine unit. In the Fuente Alegre 3 (FA-3) section, four samples have been studied and two of them have yielded conodonts. Sample 05A-93 has released *Gnathodus bilineatus bilineatus (*Roundy, 1926), and sample 05A-90 has released also *Lochriea commutata* (Branson and Mehl, 1941), *Gnathodus praebilineatus* (Belka, 1985) and *Gnathodus joseramoni* Sanz-López, Blanco-Ferrera and García López, 2004. This association indicates the same age interval than that of the other sections.

Gnathodus bilineatus bilineatus is characterised by Pa elements with nodes or ridges in the outer platform, some of them tend to alineate parallel to the carina. The inner platform presents a long parapet (Fig. II.1.26.8). However, Pa elements of *Gnathodus joseramoni* has a shorter parapet, which can take up two thirds of the inner platform length. The outer platform shows nodes that can be aligned forming concentric rows (Fig. II.1.26.1, 2, 7 and 9). *Gnathodus praebilineatus* has an outer

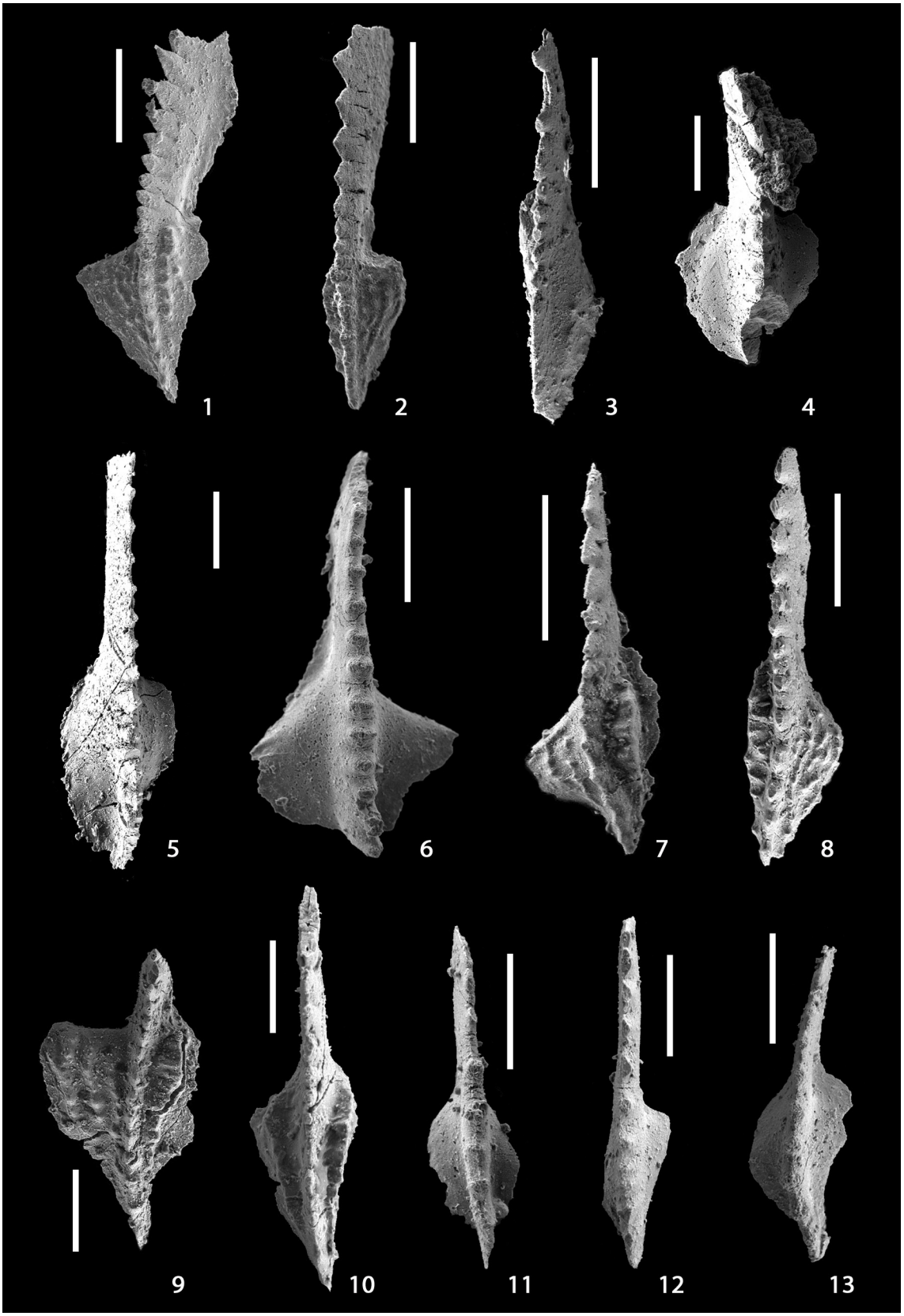


Figure II.1.26. Lower Carboniferous conodonts of the Lower Malaguide Unit (Pedrero 1: photos 1, 2, 6; Pedrero 2: photos 3- 5); Fuente Alegre 1: photos 7, 11, 12; Fuente Alegre 2: photo 8; and Fuente Alegre 3: photos 10,13). All are Pa elements oral view. Scale bars are 200 µm.

***1**, Gnathodus joseramoni; 07P-4-1. **2**, Gnathodus joseramoni; 07P-3-2. **3**, Lochriea commutata; 05A-95-4. **4**, Lochriea commutata; 05A-95-7. **5**, Lochriea commutata; 05A-94-20. **6**, Lochriea commutata; 07P-5-1. **7**, Gnathodus joseramoni; 05A-71-1. **8**, Gnathodus bilineatus bilineatus; 05A-89-12. **9**, Gnathodus joseramoni; 05A-90-6. **10**, Gnathodus praebilineatus; 05A-90-5. **11**, Lochriea commutata; 05A-71-3. **12**, Lochriea commutata; 05A-72-1. **13**, Lochriea commutata; 05A-90-1.*

platform ornamented by randomly scattered nodes (Belka, 1985). This species differs from *Gnathodus bilineatus* by having a narrower outer platform and the arrangement of the nodes.

Lochriea commutata is distinguished by Pa elements with an oval-subquadrate platform without ornamentation (Higgins, 1975) (Fig. II.1.26.3-6, 11-13).

◇◇◇

II.1.4.3. Upper Malaguide Unit

Silurian and Devonian conodonts found by van den Boogaard (1965) came from outcrops that belong to the Upper Malaguide Unit, from which two new outcrops have provided Upper Devonian conodonts: Gaticos and El Piar. In addition, the Oquendos section, also belonging to this unit, has also provided some Frasnian and Fammenian conodonts (Rodríguez-Cañero, pers. com., unpublished results).

In El Piar section several limestone pebbles from the Retamares conglomerate were sampled

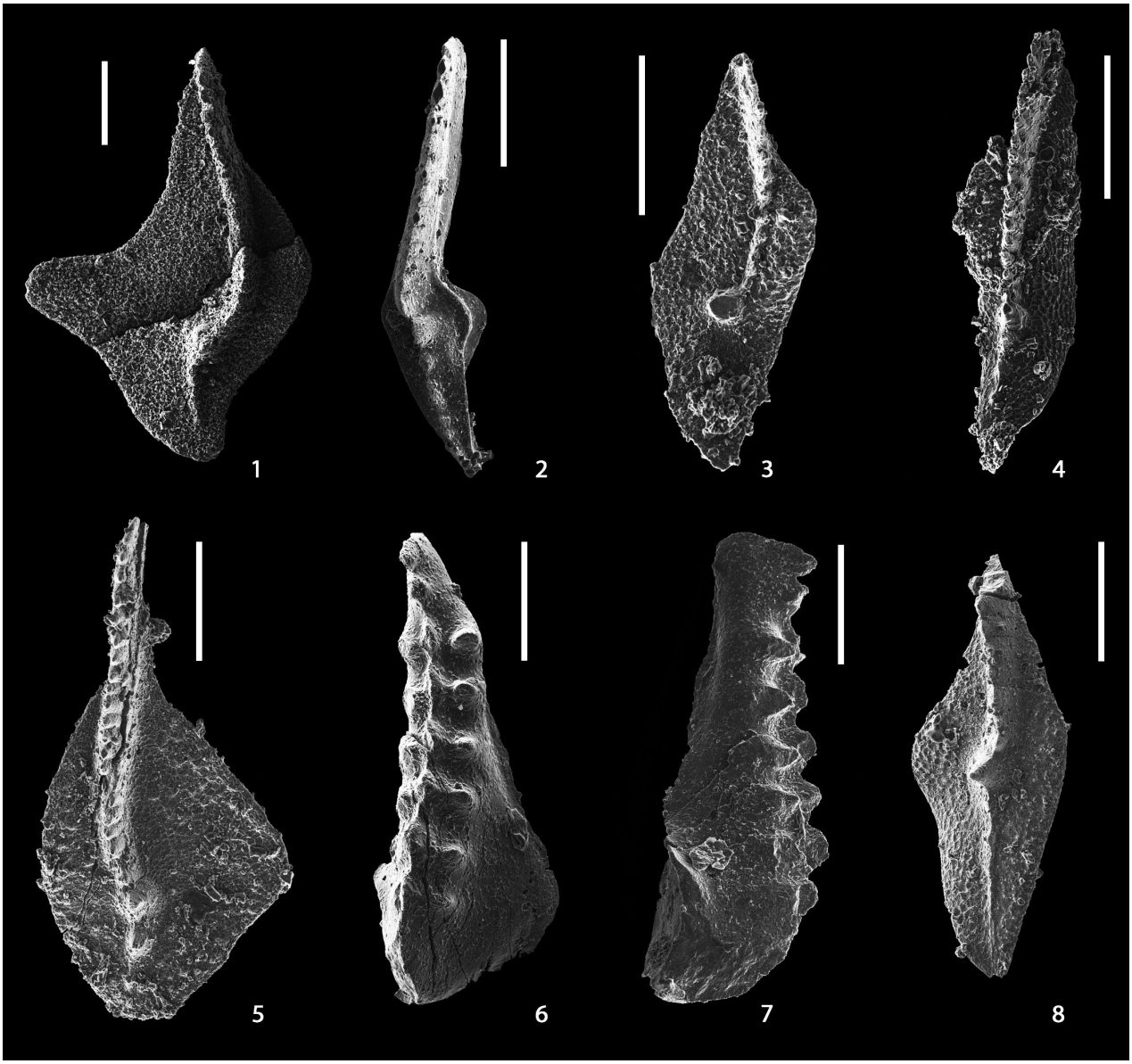


Figure II.1.27. Upper Devonian conodonts of the Upper Malaguide unit (El Piar: photos 1-2; Gaticos: photos 3-8). All are Pa elements. Scale bars are 200 µm.
***1**, Palmatolepis subperlobata; 06P-87-1. **2**, Palmatolepis gracilis gracilis; 06P-87-3. **3**, Palmatolepis glabra prima; 05A-117-1. **4**, Palmatolepis glabra prima; 05A-117-2. **5**, Palmatolepis delicatula delicatula; 05A-117-4. **6**, Icriodus symmetricus; 05A-117-6 oral view. **7**, Icriodus symmetricus; 05A-117-6 lateral view. **8**, Palmatolepis minuta minuta; 05A-117-9.*

(Fig. II.1.24). Sample 06P-87 has yielded *Palmatolepis subperlobata* Branson and Mehl, 1934, *Palmatolepis gracilis gracilis* Branson and Mehl, 1934 and Palmatolepis *minuta minuta* Branson and Mehl, 1934. These species existed during Late Devonian time but in different intervals within the Famennian stage.

Pa elements of *Palmatolepis subperlobata* show a lateral lobe well developed and a sigmoidal carina and blade (Ji, 1989). Platform is bent from the central node to the posterior end (FIg. II.1.27.1). *Palmatolepis gracilis gracilis* is represented by Pa elements with a relatively short, narrow platform and a keel that is deflected laterally below the central node (Ziegler, 1977) (Fig. II.1.27.2).

In the Gaticos section (Fig. II.1.21), sample 05A-117, which was taken from a breccia bed underlying black cherts assigned to the Falcoña Fm., has yielded Upper Devonian conodonts. Due to cementation and pebble to granule grain size, it was impossible to sample each limestone clast independently, and several different clasts were included. Therefore, conodonts indicating diverse ages were found, although all are Upper Devonian. Unfortunately, the underlying limestone has not released conodonts. *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis gracilis gracilis* Branson and Mehl, 1934 and *Palmatolepis delicatula delicatula* Branson and Mehl, 1934, which are typical Famennian species, represent most of the fauna in sample 05A-117.

Palmatolepis glabra prima has a slender platform and is characterised by a rounded, slightly convex, inner platform margin parapet (Ziegler, 1977) (Fig. II.1.27. 3 and 4). The platform outline of *Palmatolepis minuta minuta* is varied and includes forms with or without lateral lobe, but always presents a restricted anterior platform that broads in the region of the central node (Olivieri, 1970) (Fig. II.1.27.8). *Palmatolepis minuta* is similar to *Palmatolepis gracilis* in platform reduction. However, it is distinguished by not having the lateral deflection of the keel below the central node that is typical of *Palmatolepis gracilis*. In *Palmatolepis delicatula delicatula*, the anterior outer margin is characteristically straight and reaches the blade at an acute angle (Ziegler and Sandberg, 1990). The outer lobe is slightly pronounced; for this reason, the platform has an almost triangular shape (Fig. II.1.27.5).

Sample 05A-117 has also released Pa elements of *Icriodus symmetricus* Branson and Mehl, 1934 that indicates a Frasnian age. This *Icriodus* species is characterised by Pa elements that present a narrow platform that is slightly deflected inwards (Fig. II.1.27.6). The middle row of denticles is higher than the lateral rows, particularly in the posterior region, and the middle row denticles are fused (Ziegler, 1975) (Fig. II.1.27.7).

In spite of being, clearly, a mixed fauna, and due to the stratigraphic position below the black cherts that probably belongs to the Falcoña Fm., a latest Devonian age could be inferred for this breccia.

Sections	González 1		González 2			González 3		Gaticos	El Piar	
Samples	05A-108	05A-110	06P-89	06P-90	06P-91	05A-113	05A-114	05A-117	06P-87	06P-88
dissolved amount (kg)	1.000	0.900	0.450	0.400	0.475	0.550	0.500	0.175	0.075	0.075
Declinognathodus noduliferous							X			
Icriodus symmetricus								X		
Idiognathoides corrugatus							X			
Idiognathoides sinuatus						X	X			
Gnathodus bilineatus bilineatus				X						
Palmatolepis glabra prima								X		
Palmatolepis gracilis gracilis								X	X	
Palmatolepis minuta minuta								X	X	
Palmatolepis subperlobata									X	
Polygnathus cf. *costatus*		X								
Icriodus sp./fragments		11						1		
Idiognathoides sp./fragments										
Palmatolepis sp./fragments								17	10	
Polygnathus sp./fragments		15						2	8	
Fragments/ramiforms	3	29	2	5	1	6	59	27	33	4

Sections	Fuente Alegre 1		Fuente Alegre 2	Fuente Alegre 3			Pedrero 1		
Samples	05A-71	05A-72	05A-89	05A-90	05A-91	05A-93	05A-94	05A-95	05A-96
dissolved amount (kg)	1.100	1.125	1.500	1.300	0.400	1.325	0.775	0.925	0.850
Gnathodus bilineatus bilineatus			X	X		X	X	X	
Gnathodus joseramoni	X	X	X	X			X		X
Gnathodus praebilineatus				X					
Lochriea commutata	X	X	X	X			X	X	
Fragments/ramiforms	17	11	19	7	2	6	21	9	26

Sections	Pedrero 2								
Samples	07P-1	07P-2	07P-3	07P-4	07P-5	07P-6	07P-7	07P-8	07P-9
dissolved amount (kg)	0.850	0.800	0.825	0.900	0.925	0.875	0.900	0.875	0.900
Gnathodus bilineatus bilineatus	X		X	X	X			X	X
Gnathodus joseramoni		X	X	X					
Gnathodus praebilineatus		X		X					X
Lochriea commutata		X			X			X	
Fragments/ramiforms	7	49	50	38	49	17	11	25	32

Table II.1. Conodont elements abundance and distribution in the studied sections in the Vélez Rubio-Lorca Corridor.

II.2.1. Geological setting and general tectonic subdivision of the Malaguide Complex in the area

The Cogollos Vega Zone was defined by Blumenthal and Fallot (1935) and comprises a narrow WSW-ENE topographic depression, to the south of Sierra Arana, between the Sierra de Cogollos and the Sierra de Huétor, and from the Granada Basin to the Guadix Basin (Fig. II.2.1.A). This area includes the widest outcrop of the Malaguide Complex in the Granada province.

South of Sierra Arana, in the Cogollos Vega Zone, Blumenthal (1928) was the first author who recognised the presence of the Betic of Málaga -defined by himself (Blumenthal, 1927) in the western Betic Cordillera, and which later was named Malaguide Complex by Durand-Delga (1968).

Regarding to the structure (Fig. II.2.1.B) of the Malaguide Complex in the Cogollos Vega Zone, it tectonically overlies the uppermost units of the Alpujarride Complex by means of a N-dipping contact (Durand-Delga and Fontboté, 1960). A subvertical fault (Almuéjar fault of Balanyà, 1984) separates the Cogollos Vega Zone from the Frontal Units of the Internal Domain in this area (Martín-Algarra *et al.*, 2004d). The Frontal Units (also named Dorsale bétique by Durand-Delga and Foucault, 1967) crop out immediately to the north of the Malaguide Complex, where they form different

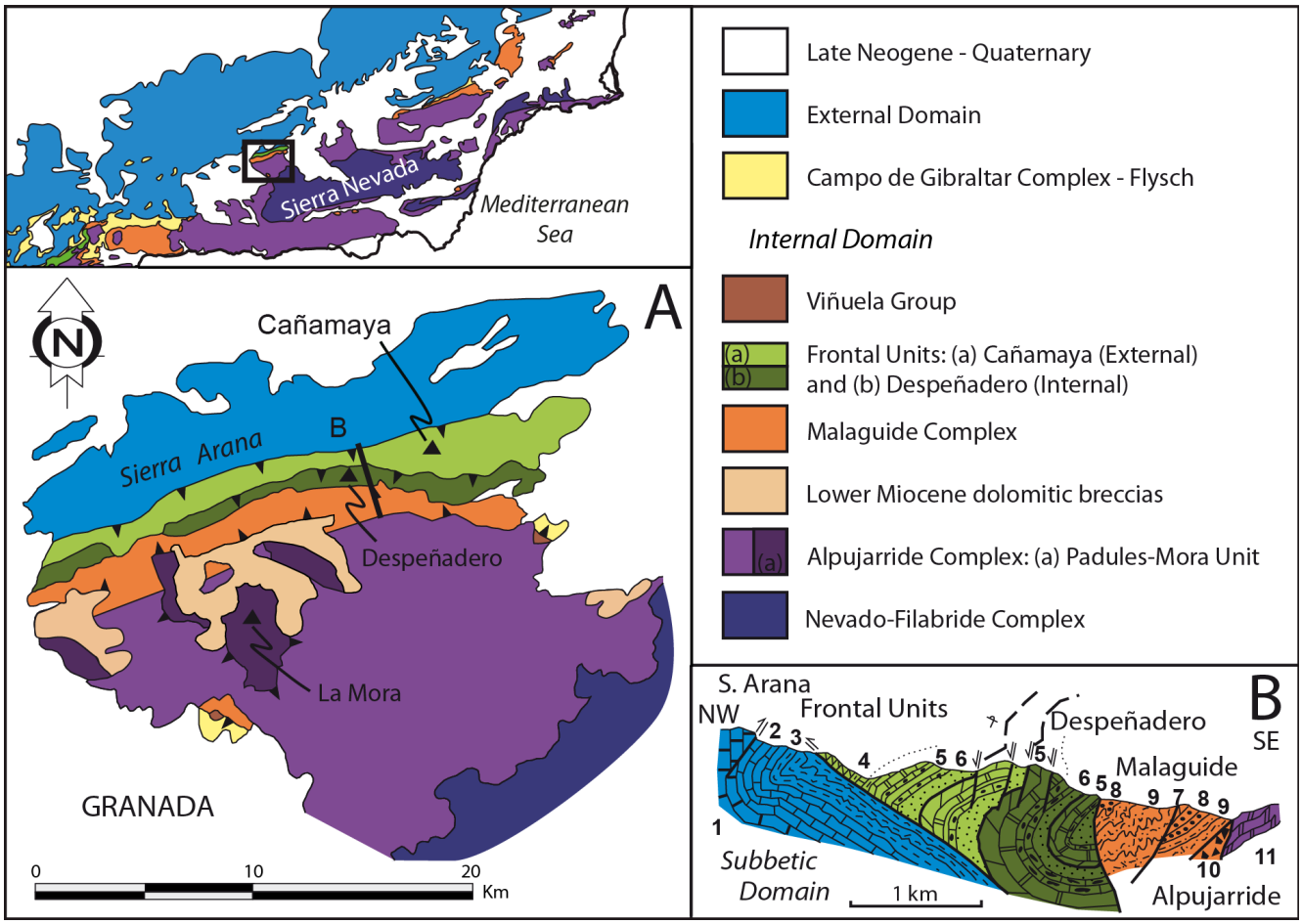


Figure II.2.1. ***A.*** *The Cogollos Vega Zone geological sketch.* ***B.*** *Synthetic geological section.*

thrust-bounded units with frequently vertical-to-reversed stratigraphic successions (Fig. II.2.2.).

In the Cogollos Vega Zone Durand-Delga and Foucault (1967), and later Foucault (1976), García-Dueñas and Navarro-Vilá (1976) and Navarro-Vilá and García-Dueñas (1979), identified several Alpine Malaguide units, which are mainly made up of olive-green Paleozoic slates and sandstones, and of red Triassic rocks. The successions are organised into two to four Alpine tectonic units depending on the location of the section. These units have a limited lateral continuity, and García-Dueñas and Navarro-Vilá (1976), in particular, identified three main imbricated Alpine Malaguide units, respectively named by them *1*, *2* and *3* from bottom to top (and from south to north, although they mention that they were locally modified by later tectonics). Their lowermost unit, *Unit 1*, would be made up of Permian?-Triassic Pseudoverrucano facies that belong to the Saladilla Fm defined in the Vélez Rubio area. Their *Units 2* and *3* include Paleozoic and Triassic deposits. Nevertheless, the distribution of the Malaguide units proposed by the same authors in the geological map later published by themselves in 1979 (Navarro-Vilá and García-Dueñas, 1979) does not fit with those previously mapped.

More recently, Sanz de Galdeano *et al*. (1995a, b) studied the general structure of the Cogollos Vega Zone. They also revised the Alpujarride and Malaguide tectonic units proposed by Foucault (1976), by García-Dueñas and Navarro-Vilá (1976) and by Navarro-Vilá and García-Dueñas (1979). In the northern sector of the Cogollos Vega Zone, Sanz de Galdeano *et al*. (1995a, b) described several S-vergent tectonic slices, showing typical Malaguide successions to the north, and overlying units showing a stratigraphy and tectonic structure with progressively transitional features to those of the Alpujarride units, which they called *Transitional Units*. Actually, when these units occur in lower tectonic levels, both the Paleozoic basement and the Permian?-Triassic deposits show a more intense deformation and metamorphic recrystallization, reaching low-grade metamorphic conditions and developing a better and better defined schistosity. In particular, the Garay Unit is described by Sanz de Galdeano *et al*. (1995a, b) as an intermediate unit with features between those of Malaguide and Alpujarride successions. They considered it to be more related to the Alpujarride Complex although they pointed out the resemblance of their Paleozoic basement with that of the Malaguide Complex.

According to our own observations, the Alpujarride-Malaguide transitional units above mentioned by Sanz de Galdeano *et al*. (1995a, b), are equivalent to the Epimetamorphic Malaguide Unit described in the Vélez Rubio-Lorca area (see section II.1). Their most typical outcrops are located between the Venta del Molinillo and the Alto de las Negrillas, which is mainly formed by Triassic rocks, although a narrow belt of Paleozoic terrains with strong Malaguide affinity is also present to the south, mainly outcropping to the right-hand bank of the Fardes River (Fig. II.2.2). Another outcrop appears from the El Púlpito tectonic window towards the east, up to the Sierra de la Yedra area (Fig. II.2.2).

In addition, as in the Vélez Rubio area, the Malaguide imbrications can be grouped in two main Alpine tectonic units showing slightly different Paleozoic successions. The Lower Malaguide Unit is the largest one in outcrop, bears the *Calizas alabeadas* facies of the Santi Petri Fm. at its stratigraphic-tectonic base, followed, in normal stratigraphic succession, by strongly folded black cherts and limestones of the Falcoña Fm. or, when the later has been eroded or tectonically laminated, by the immature turbiditic siliciclastic deposits with Culm facies of the Almogía Fm. In addition, the Lower Malaguide Unit is also related to large outcrops of the Triassic Saladilla Fm., but, as in the Vélez Rubio-Lorca area, no Jurassic or younger rocks have been found related to them. Nevertheless, in the Upper Malaguide Unit, which seems to be subdivided in several smaller slices, the Santi Petri Fm. is absent, and the Devonian beds are formed by slates and fine grained sandstones including, sporadically, laterally discontinuous lenses of dacryoconarids-bearing limestones. In addition, above the widespread pelites and sandstones and associated conglomerates of the Almogía Fm., the Marbella Fm. conglomerates, including cobble- to boulder-sized clasts of shallow water Carboniferous limestones and of granite and gneiss (Heereken, 2003), are exclusively related to this Upper Malaguide Unit. This unit also included Triassic redbeds and Jurassic to Miocene deposits. However, the latter are badly characterised in the Cogollos Vega Zone due to the very strong tectonization and poor outcrop conditions in most areas.

II.2.2. Previous studies and general features of the Paleozoic stratigraphy

According to García-Dueñas and Navarro-Vilá (1976) and Navarro-Vilá and García-Dueñas (1979), the Paleozoic successions of the Malaguide Complex in the Cogollos Vega Zone are characterised by the presence of the following facies associations, from bottom to top, stratigraphically:

a) Glossy slates, quartzites, conglomerates with stretched pebbles and, to the top, several limestone levels. The authors tentatively assign a Silurian age to this facies association without any biostratigraphical base. According to their macroscopical features and to their distribution in map view, I correlate these lithofacies to the Morales Fm.

b) Pelites, greywackes and conglomerates, with intercalated *Calizas alabeadas* facies, which correspond to the Santi Petri Fm.

c) Pelites, conglomerates, calcareous sandstones and greywacke intercalations with undeterminable plant remains. This interval can be correlated to the Culm deposits of the Almogía Fm.

This succession mainly corresponds to that of the Lower Malaguide Unit, which constitutes the southernmost belt of the Malaguide outcrops in the Cogollos Vega Zone, apart from those outcrops that were affected by the Alpine epimetamophism.

García-Dueñas and Navarro-Vilá (1976) and Navarro-Vilá and García-Dueñas (1979) interpreted that the topmost part of the facies association c above mentioned, corresponding to the Almogía Fm., would be constituted by the heterometric and polymict conglomerate of Marbella, i.e., the Marbella Fm. Nevertheless, as stated above, the latter are exclusively present in outcrops related to the Upper Malaguide Unit and can be found further north.

Previously, Foucault and Paquet (1970, 1971), in the vicinity of the Cortijo de Florencia, identified strongly folded, black chert horizons (lydites) unconformably overlain by reddish Triassic sandstones and conglomerates. These black cherts can be assigned to the radiolaritic member of the Falcoña Fm., and can be found here and there in many small and strongly tectonized outcrops along the Cogollos Vega Zone, mainly, but not exclusively as we will see later, in relation to the Lower Malaguide Unit. These authors also indicated the existence, to the south of the Majalijar, of a small lens of tentaculites limestone, which we were unable to find yet probably due to its small size.

Herbig (1984) studied microfacies and fossil content of three outcrops of the Marbella Fm. in the Cogollos Vega Zone, all three belonging to the Upper Malaguide Unit (Fig. II.2.2). He concluded the Marbella Fm. corresponds to debris flow deposits. But he could not establish the transport direction of the conglomerates and thus neither the position of the continental slope. Notwithstanding this, he stated that a NW-SE bipolarity prevails in the Cogollos Vega Zone. Heekeren (2003) also sampled and studied carefully the limestone pebbles within this formation. He identified several shallow and pelagic marine facies, some of them bearing Devonian and Carboniferous fossils.

Good outcrops of Malaguide rocks in the Cogollos Vega Zone are scarce due to dense forest cover and soils, and to strong tectonization. Nonetheless, numerous observations can be made in isolated small outcrops along trails, paths and small gullies. When they have been done in outcrops relatively close to each other, it is possible to reconstruct a general stratigraphic column of the area. The development of the previously defined classical Paleozoic formations of the Malaguide Complex is very different from site to site. The most complete successions and widest outcrops observed in the Cogollos Vega Zone belong to the Lower Malaguide Unit. However, the Morales Fm. has not been clearly recognised in this unit or, rather, it could not be clearly separated from pelitic-arenitic facies related to the Santi Petri Fm. The Marbella Fm. occurs exclusively in the Upper Malaguide Unit.

For this dissertation, in addition to scattered field observations, the Paleozoic stratigraphy has been reconstructed in three areas of the Cogollos Vega Zone, which approximately correspond to N-S general cross sections of the Malaguide Complex, from west to east: i) one from the Cortijo de Florencia area to the south of the Majalijar; ii) another in the Las Mimbres area (to the east of the Majalijar); and, finally, iii) to the west of the Venta del Molinillo, from the Alto de las Negrillas to the Cortijo de la Mina, in particular in the vicinity of the cortijos del Mojino and de la Solana. In addition, some observations were also made from the Venta del Molinillo to the Cortijo de la Ermita, close to the Despeñadero. All of them are located in the central sector of the Cogollos Vega Zone (Fig. II.2.2), and crosscut the three groups of Alpine Malaguide tectonic units above mentioned.

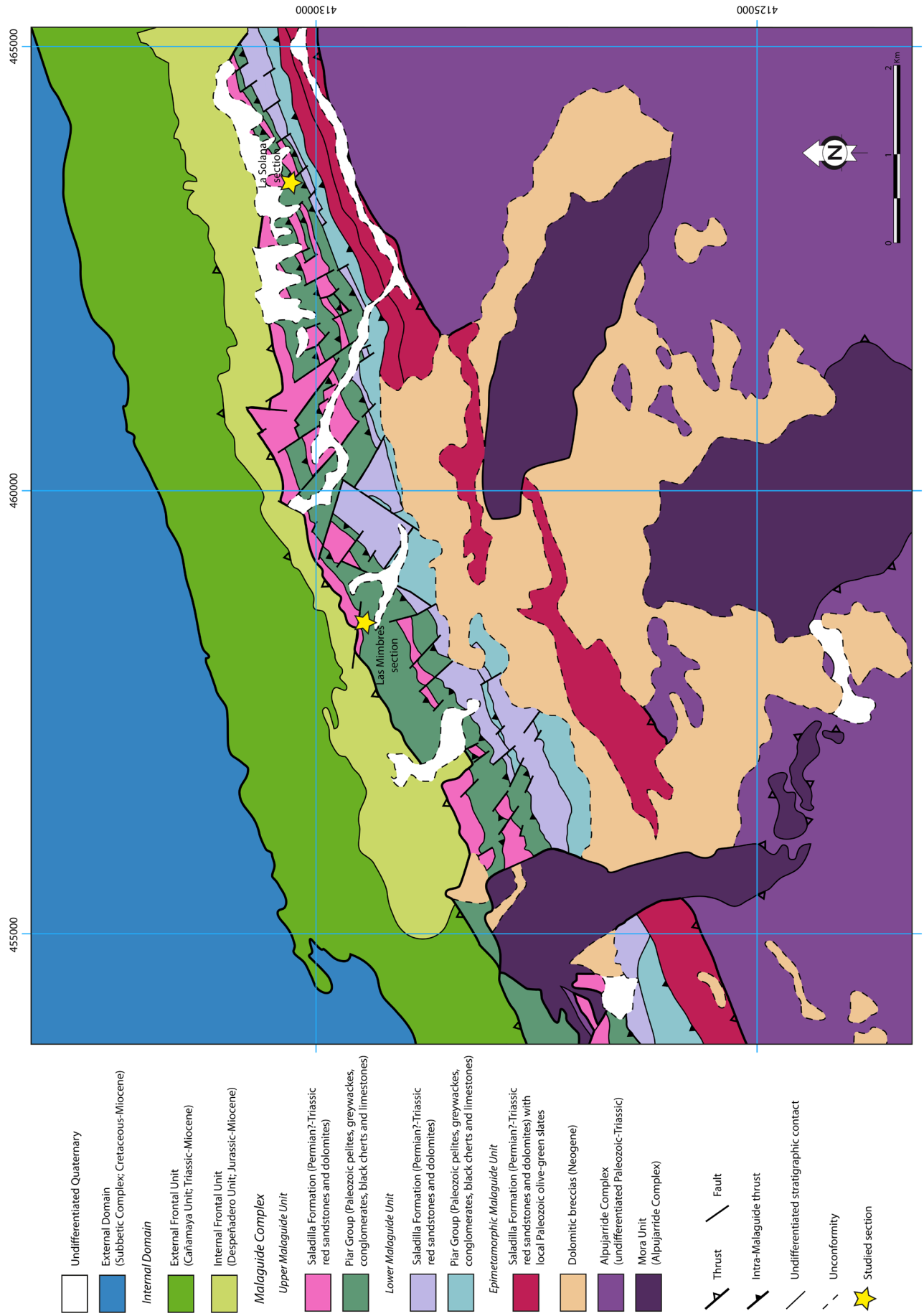


Figure II.2.2. Malaguide tectonic units and location of the studied sections in the Cogollos Vega Zone.

II.2.3. Studied sections

A geological section of the Malaguide terrains from the Florencia area to the Majalijar reveals, in addition to the local presence of Malaguide-like Triassic epimetamorphic red pelites between distinctive Alpujarride and Malaguide terrains, typical Malaguide Paleozoic and Triassic successions as mentioned above. Most Paleozoic lithotypes in this area are, however, formed by pelites and greywackes that mainly belong to the Almogía Fm., either when associated with the Lower or with the Upper Malaguide Unit, with only local presence of black cherts. These lydites, attributed to the lower member of the Falcoña Fm., have been found in the Florencia area and to the SW of las Mimbres as a part of the Lower Malaguide Unit, overlying brown slates and greywackes including bluish turbiditic limestone beds corresponding to the *Calizas alabeadas* facies of the Santi Petri Fm.

In a geological cross section from the El Calabozo towards the north, the Epimetamorphic Malaguide Unit, widely outcropping towards the west and east, is lacking, and the Lower Malaguide Unit lies directly on Alpujarride rocks and associated rhauwacke-like carbonate breccias overlying them (Fig. II.2.2). South of the road from the Alfaguara to Prado Negro, most Paleozoic terrains belong to the Lower Malaguide Unit, which shows its typical features and is formed by the Santi Petri, Falcoña and Almogía formations in normal stratigraphic succession dipping to the north. These Paleozoic formations are covered by an important outcrop of strongly tectonized Triassic terrains (Saladilla Fm.). North of the road, and bound from Triassic terrains located to the south by a subvertical fault, most Paleozoic terrains can be included within the Upper Malaguide Unit, and, at least in part of this area, they are stratigraphically upside down (see below).

In the picnic area of Las Mimbres (Fig. II.2.2) the Paleozoic rocks, dipping to the north, can be observed at the Fuente de los Potros, along the forestal track to Prado Negro, where they are mainly made of dark-green slates with thin to very thin greywacke intercalations (Fig. II.2.4.e). The thickest greywacke beds locally preserve some sedimentary structures, and include graded conglomeratic horizons showing channeled structures at the base of stratigraphically reversed strata (Fig. II.2.4.f). The conglomerate beds are normally graded, however, when restituted to their original subhorizontal position, and are made up of clasts of pebble and granule size. These clasts are mainly made of fine-grained greywacke and white quartzite, but also of black lydite. These rocks are totally lacking fossils (except for some unclassifiable plant remains) and correspond to the Culm facies, being then equivalent to the Almogía Fm. (including representatives of the Retamares conglomerate member).

To the north, along the path from Las Mimbres picnic area to the Acequia Fardes (via A-92 highway, exit 264), in the surroundings of the Fuente de los Potros, a discontinuous limestone lens dipping to the north crops out (Las Mimbres section, Fig. II.2.3). This lens is made up of several limestone beds separated by slate partings (Fig. II.2.4. b, c, d). It appears isolated between dark green slates (Fig. II.2.4.a) although black chert blocks occur in the neighbourhoods. Two samples were taken from the limestone beds and released Lower Devonian conodonts (see section II.2.3. for biostratigraphic details). This biostratigraphic results have been published during the course of this dissertation (Navas-Parejo *et al*., 2011, see appendix I). Since the Almogía Fm. appears to the south, and according to both sedimentological field observations and biostratigraphic data, the Paleozoic succession in Las Mimbres area is stratigraphically reversed. Close to the contact with the Internal Frontal Units to the north of Las Mimbres, Herbig (1984) studied another outcrop of the Marbella Fm.

Just to the south of the Majalijar, near to the stratigraphic top of the Upper Malaguide Unit, there exists one important outcrop of the Marbella conglomerate Fm. along the Acequia del Fardes. An equivalent succession can be observed in the path from la Venta del Molinillo to the Cortijo de la Ermita and to the Cortijo del Despeñadero, where there exists another important outcrop of the Marbella Conglomerate Fm. In both areas, and contrary to what can be observed in Las Mimbres area, located between them, the successions of both the Lower and the Upper Malaguide Units seems to be in normal stratigraphic position.

In the Alto de las Negrillas (A-92 highway, exit 268), typical Triassic Malaguide Pseudoverrucano deposits widely crop out, dipping to the north (Fig. II.2.5.B). In the trench of the secondary road from the exit 268 to the Venta del Tejarillo and the Venta del Molinillo, black brecciated dolomites overlying glossy purplish slates crop out. They show slight metamorphism, higher than in the other Malaguide units (see also Sanz de Galdeano *et al*. 1995). Because

of this and despite Paleozoic rocks do not crop out in this Malaguide Unit, I assign these Triassic rocks to the Epimetamorphic Malaguide Unit.

From the Venta del Tejarillo to the NE, along a private track parallel to the watercourse, and over the Epimetamorphic Malaguide Unit, the Paleozoic succession of the Lower Malaguide Unit occurs (Fig. II.2.5). This Paleozoic succession starts with about 30 m of green-olive slates and muscovite-rich greywackes, followed by about 50 m of slates and greywackes with platy *Calizas alabeadas* facies (Fig. II.2.6.a and b). The latter are tectonically laminated and do not show here the typical warps or *alabeos*. These Paleozoic rocks are unconformably covered by a few tens of metres of Triassic red pelites, sandstones and conglomerates.

Overlying the Triassic rocks, the Upper Malaguide Unit appears (figs. II.2.5.C, II.2.6.c and d). The Paleozoic succession is made up of about 40 metres of pelites with greywacke intercalations (Fig. II.2.6.e), unconformably covered by Triassic reddish pelites and sandstones with greenish basic volcanic rocks and strongly brecciated tectonic lenses of Jurassic to Tertiary rocks. In the surroundings of the Cortijo el Mojino, a badly exposed Paleozoic succession that seems to be tectonically sandwiched between the Triassic rocks of this Upper Malaguide Unit, includes a metre-sized lens formed by bluish (yellowish when weathered) platy limestones bearing dacryoconarids (Fig. II.2.6.f).

The last section studied in this area has been made along the same private track, to the east, in La Solana area (Fig. II.2.5.D). There the Paleozoic succession starts tectonically overlying red Triassic sandstones, with about 15 m of Devonian green olive pelites and greywackes with very thin beds of yellowish platy limestones bearing dacryoconarids (Fig. II.2.7.a). The succession is gently dipping to the north and continues with about 10 m of strongly folded black cherts and a few metres of greenish and reddish slates topped by platy limestones with black chert nodules (Fig. II.2.7.b and c), which have yielded Visean conodonts (see section II.2.4 for biostratigraphic details), indicating a normal stratigraphic sequence. It continues in apparently conformable succession by several tens of metres of strongly folded (SE vergent kinked bands) light brown-silvery slates (Fig. II.2.7.d). The latter rocks are followed northwards by a strongly faulted lens, around 2 m thick, of subvertical bluish limestone beds that have yielded some Upper Devonian conodonts (see section II.2.4 for biostratigraphic details) and that are followed by a few metres of

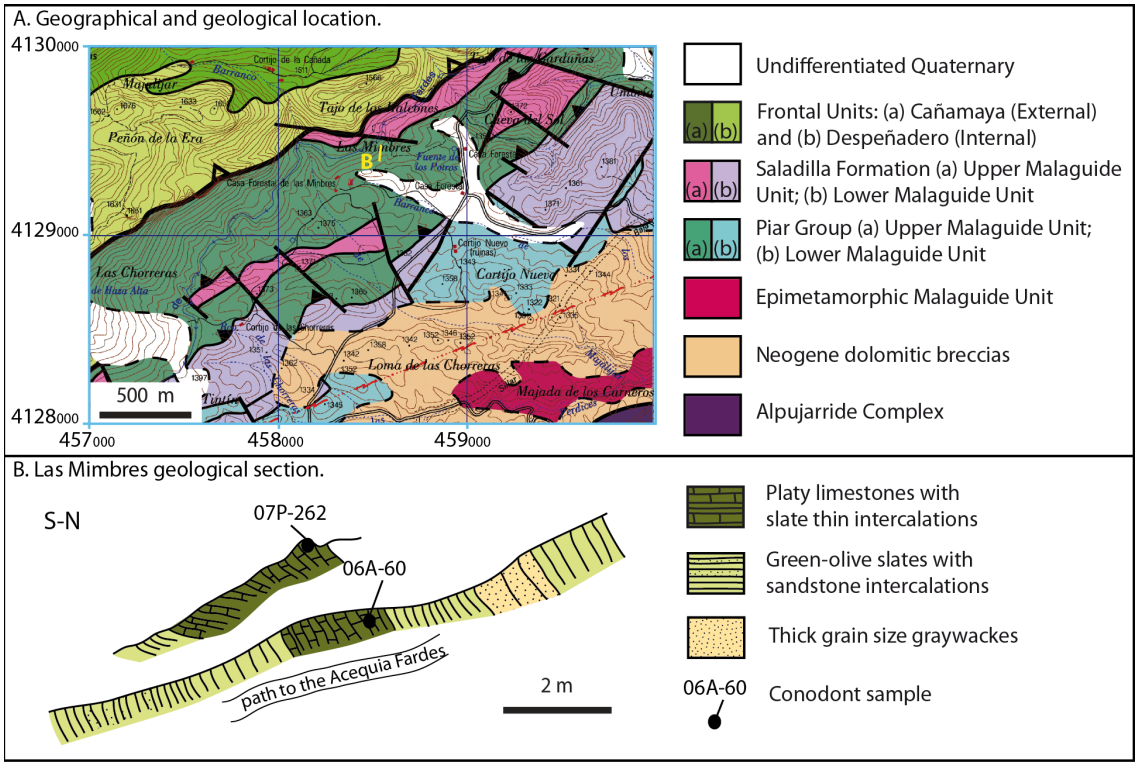


Figure II.2.3. Las Mimbres section.

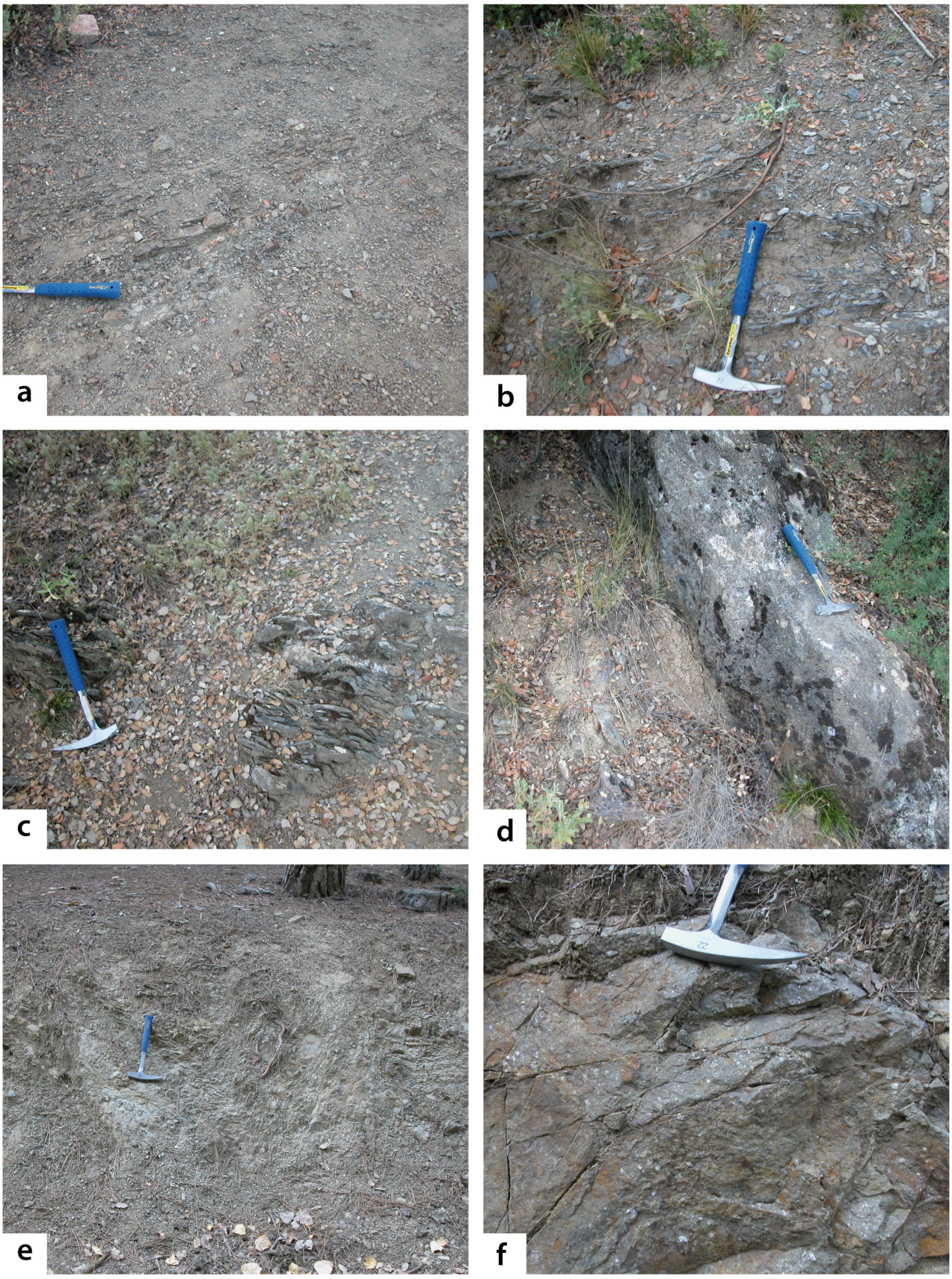


Figure II.2.4. Las Mimbres picnic area. ***a****, Green olive pelites with thin sandstone intercalations;* ***b****, and* ***c****, Platy limestones with pelitic partings, which have released Lower Devonian conodonts (sample 06A-60);* ***d****, Massive grey limestone bed with calcite veins, stratigraphically equivalent to b and c, which also yielded Lower Devonian conodonts (07P-262);* ***e****, Strongly folded slates of the Almogía Fm.;* ***f****, Retamares conglomerate of the Almogía Fm.*

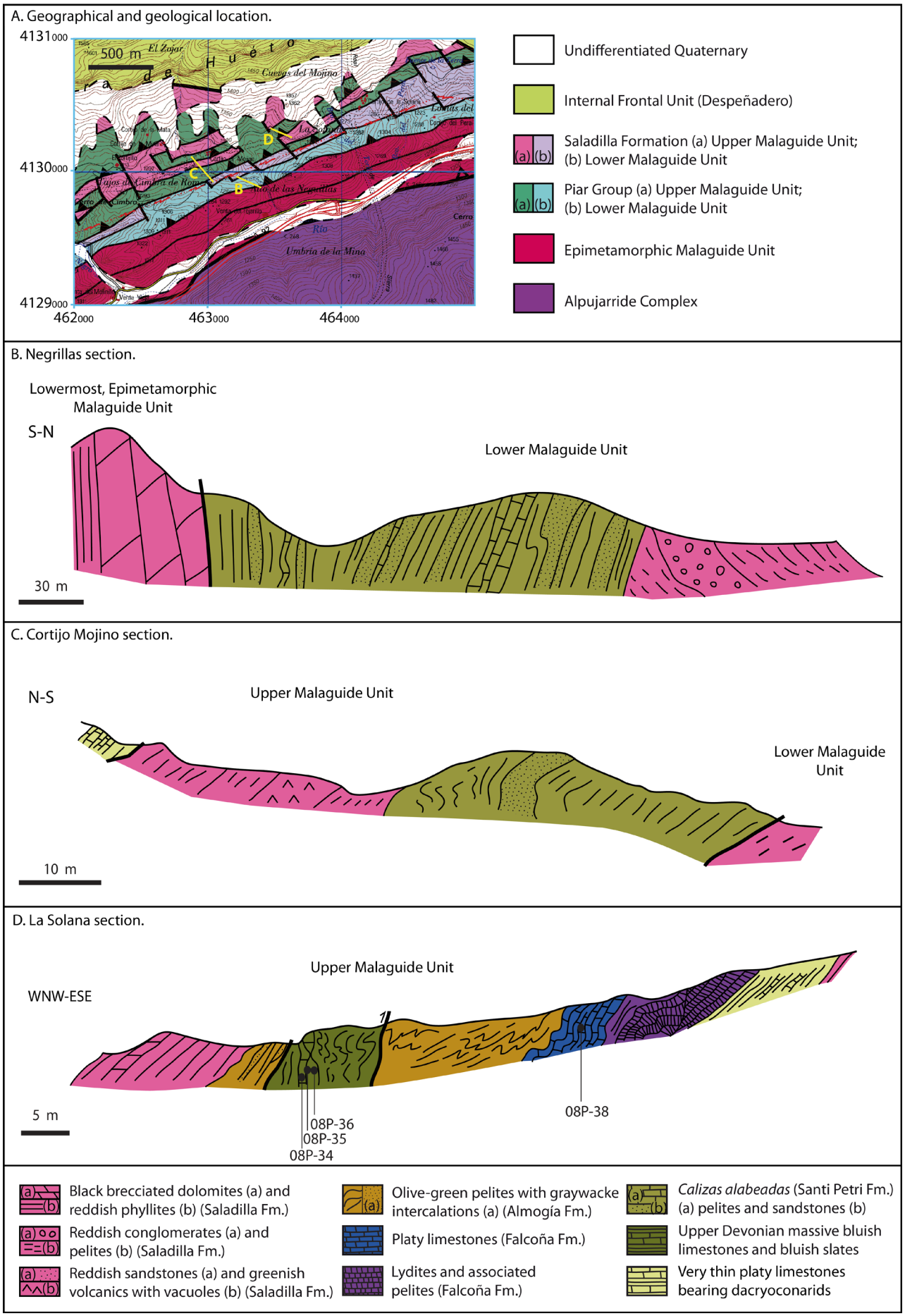


Figure II.2.5. The Negrillas, Mojino and La Solana geological cross sections.

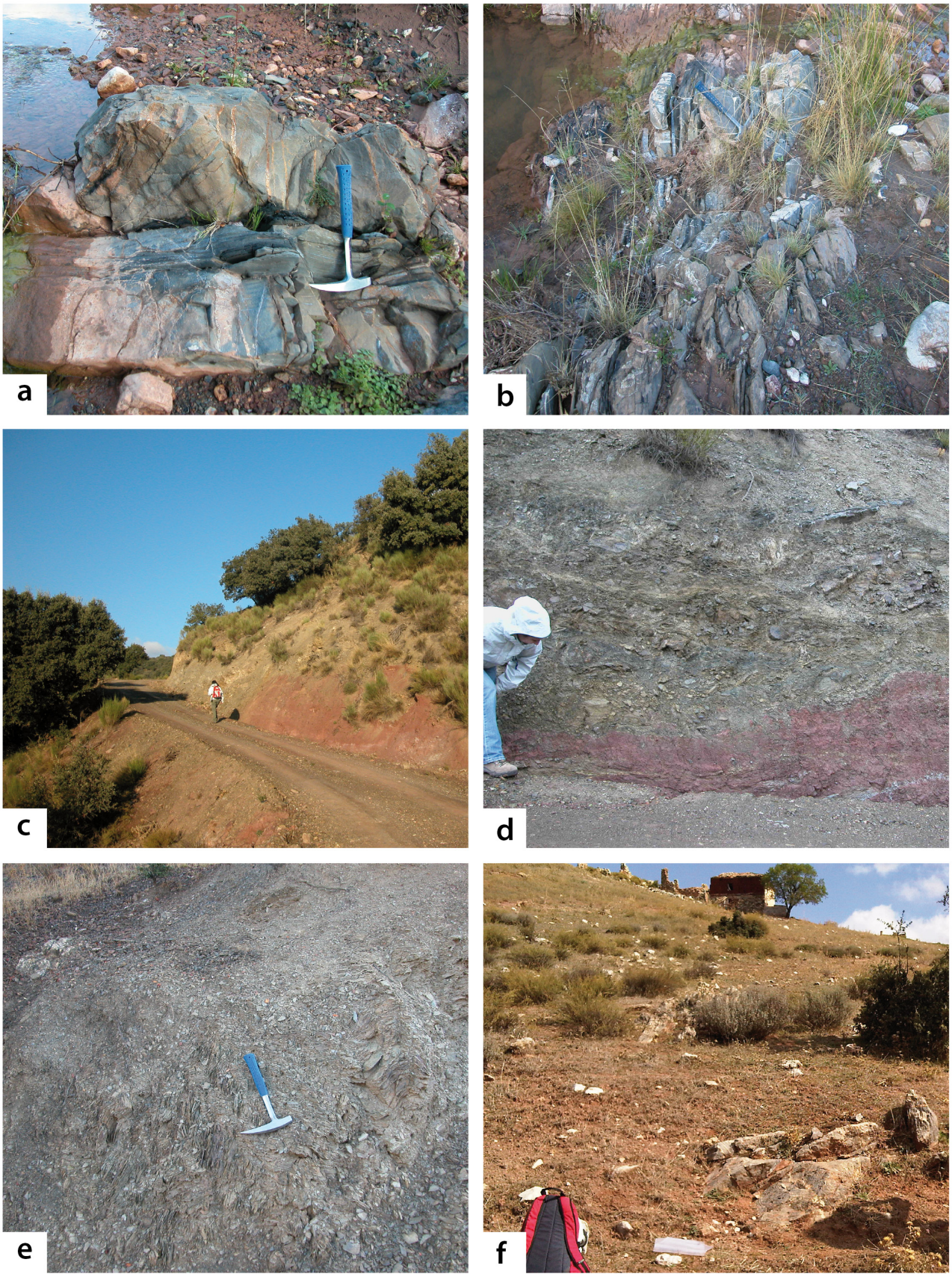


Figure II.2.6. Negrillas and Mojino sections. ***a****, and* ***b****, Thin bedded Calizas alabeadas of the Santi Petri Fm.;* ***c****, and* ***d****, Thrust between the Lower Malaguide Unit (right) and the Upper Malaguide Unit (left);* ***e****, Pelites with greywacke intercalations of the Upper Malaguide Unit;* ***f****, Isolated blocks of limestones bearing dacryoconarids near to the Cortijo Mojino (in the background).*

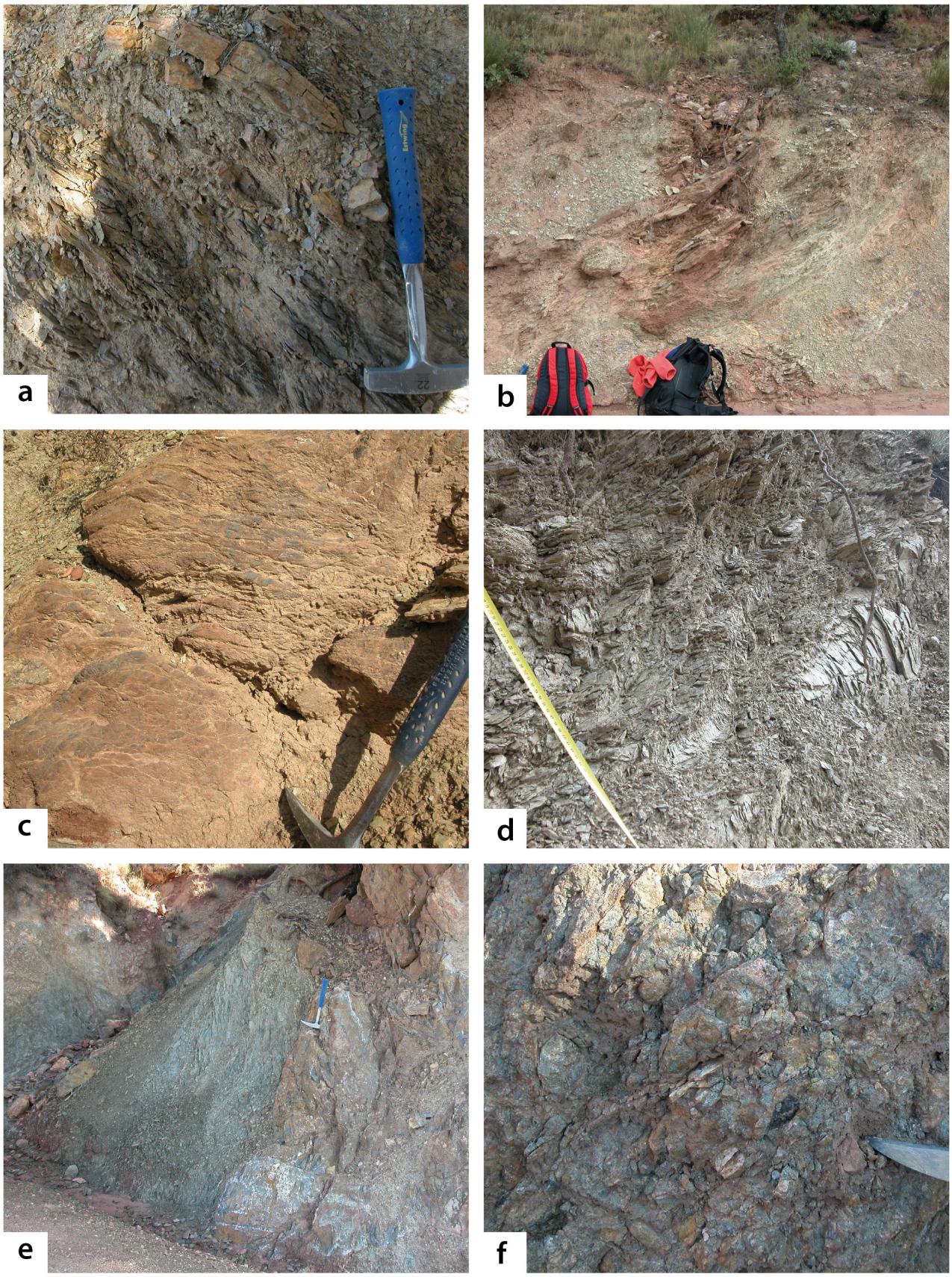


Figure II.2.7. La Solana section. ***a****, Yellowish platy limestones bering dacryoconarids;* ***b****, Lydites (right), greenish and reddish pelites, and very thin bedded limestones (Falcoña Fm.);* ***c****, Limestones of the Falcoña Fm., which have yielded one Lower Carboniferous conodont (sample 08P-38);* ***d****, Strongly folded slates of the Almogía Fm.;* ***e****, Medium bedded limestones (right), which have released Upper Devonian conodonts (samples 08P-34, 08P-35 and 08P-36) and slates (left);* ***f****, Retamares conglomerate (Almogía Fm.).*

gray pelites including some fine grained and thin-bedded sandstone beds (Fig. II.2.7.e). These beds are unconformably covered by brownish-greenish slates and greywackes that intercalate dm-thick conglomerate beds including granule- to pebble- to (rarely) cobble-sized clasts of black lydite, white quartzite, greywacke and gneiss (Fig. II.2.7.f), and showing the typical facies of the Retamares conglomerate member of the Almogía Fm.

The latter deposits are unconformably covered by a few metres of Triassic red cross-bedded sandstones followed by reddish claystones with some gypsum and by strongly tectonized Jurassic carbonate deposits.

II.2.4. Conodont biostratigraphy

A total of eleven samples for conodont analysis were taken from limestone horizons along two of the three general cross sections studied within the Cogollos Vega Zone. Positive results have been obtained only in two of the several outcrops sampled, and both belong to the Upper Malaguide Unit (Las Mimbres and La Solana sections). The total amount of sample dissolved in each case and the number and names of specimens obtained are indicated in table II.2.

In Las Mimbres section (Fig. II.2.3) six samples have been taken from the limestone levels, although only two samples have released conodonts (06A-60 and 07P-262). Both samples have yielded *Criteriognathus steinhornensis* (Ziegler, 1956), indicating an Emsian age (uppermost Lower Devonian).

Pa elements of *Criteriognathus steinhornensis* are characterised by two clearly asymmetric lateral lobes with a rounded to squared outline (Carls *et al.*, 2007) (figs. II.2.8.1-3). In addition, sample 06A-60 also yielded *Pseudooneotodus beckmanni* (Bishoff and Sannemann, 1958), which is represented by coniform elements with a curved peak with an anterior groove that looks like a hump (Lane and Ormiston, 1979) (Fig. II.2.8.4).

In La Solana section (Fig. II.2.5.D), four samples from a total of five have yielded conodonts (samples 08P-34, 08P-35, 08P-36 and 08P-38). The oldest conodont associations have been found in the bluish limestones towards the upper part of the succession (samples 08P-34, 08P-35 and 08P-36).

Sample 08P-34 released *Palmatolepis crepida* Sannemann, 1955b, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis quadrantinodosalobata* Sannemann, 1955a and *Palmatolepis tenuipunctata* Sannemann, 1955b. These species coexisted during the *crepida* Zone (Famennian; Upper Devonian).

Palmatolepis crepida is characterised by Pa elements with a drop-shaped outline and a posterior end bent upwards (Ziegler, 1974) (Fig. II.2.8.5). *Palmatolepis quadrantinodosalobata* is distinguished by Pa elements with a cluster of nodes on the inner anterior platform and a prominent outer lobe (Sandberg and Ziegler, 1973) (Fig. II.2.8.6). *Palmatolepis minuta minuta* Pa elements typically show a small, very narrow platform and a flat lateral lobe (Fig. II.2.8.7). *Palmatolepis tenuipunctata* is represented by Pa elements with an elongated platform, a small and rounded outer lobe and a posterior end bent upwards (Fig. II.2.8.8).

Sample 08P-35 yielded *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis regularis* Cooper, 1931 and *Palmatolepis glabra glabra* Ulrich and Bassler, 1926. This conodont association corresponds to the Upper and Uppermost *crepida* zones (Famennian; Upper Devonian).

Palmatolepis regularis is distinguished by Pa elements sinuously curved with subparallel margins (Metzger, 1994) (Fig. II.2.8.9 and 10).

Sample 08P-36 released *Icriodus alternatus alternatus* Branson and Mehl, 1934, and *Palmatolepis regularis* Cooper, 1931, which also belong to the crepida Zone (Famennian; Upper Devonian).

Icriodus alternatus alternatus is characterised by a straight platform where the denticles of the central row alternate with the denticles of the lateral rows, although this species is mainly distinguished by the strong lateral compression of the denticles in the central row (Sandberg *et al.*, 1992) (Fig. II.2.8.11).

Sample 08P-38 has been taken in the limestones overlying the lydites. In this sample, only one, but

very well preserved, Pa element of *Gnathodus* cf. *bilineatus bilineatus* (Roundy, 1926) has been found. This species ranges from the *bilineatus* to the *bollandensis* Zone (Lower Carboniferous). *Gnathodus bilineatus bilineatus* is distinguished by a long parapet parallel to the carina that reached the posterior tip. The outer platform has a square outline and present nodes and ridges, some of which tend to align parallel to the posterior carina. Specimen found shows these features although the nodules attached to the carina are not very well developed (Fig. II.2.8.12).

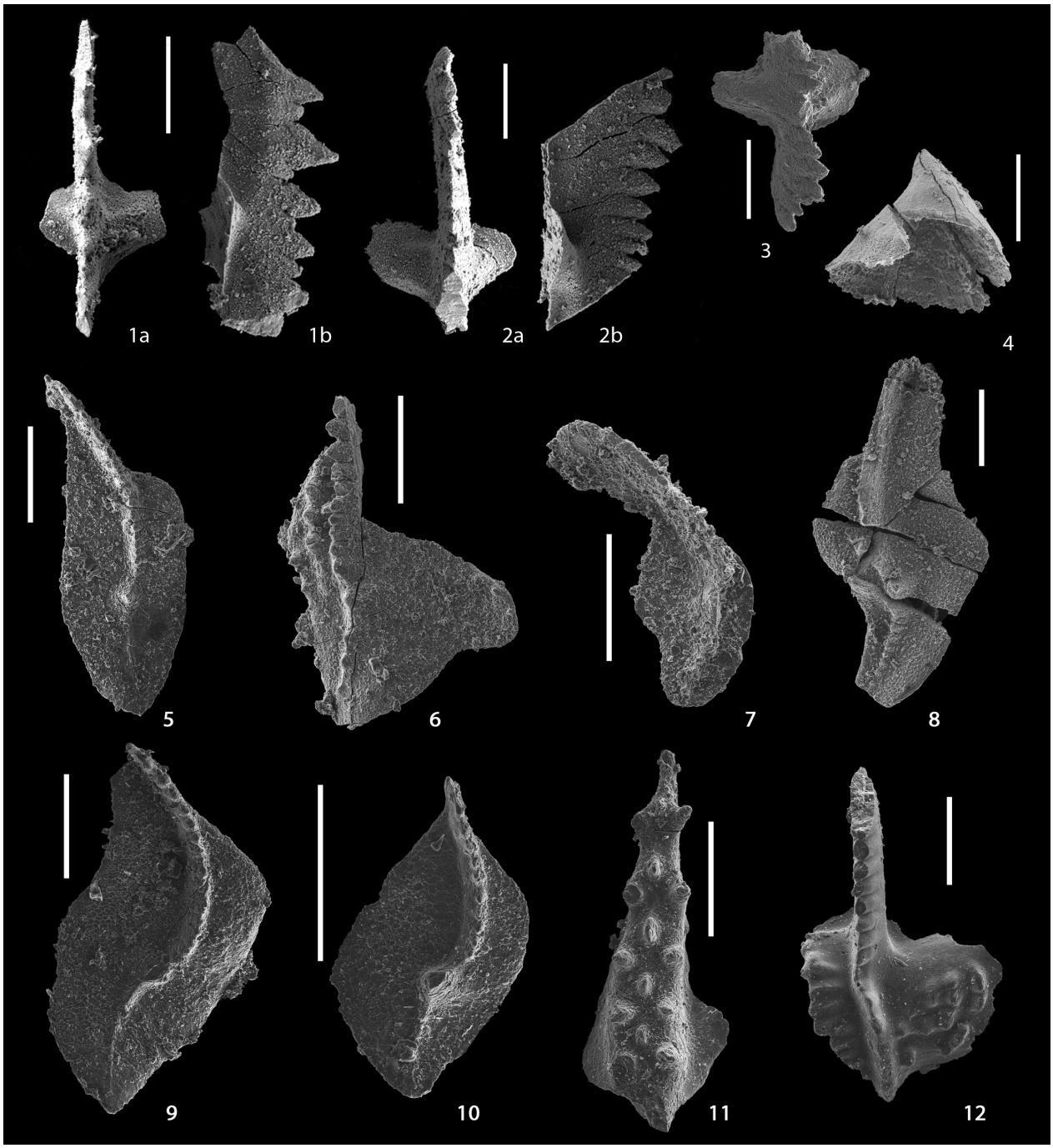


Figure II.2.8. Devonian and Carboniferous conodonts in Las Mimbres (photos 1-4) and La Solana (photos 5-12). All are Pa elements. Scale bars are 200 µm.

***1**, Criteriognathus steinhornensis; 1a, upper view; 1b, lateral view; 06A-60-4. **2**, Criteriognathus steinhornensis; 2a, upper view; 2b, lateral view; 06A-60-1. **3**, Criteriognathus steinhornensis; 07P-262-1. **4**, Pseudooneotodus beckmanni; 06A-60-5. **5**, Palmatolepis crepida; 08P-34 008. **6**, Palmatolepis quadrantinodosalobata; 08P-34 012. **7**, Palmatolepis minuta minuta; 08P-34 010. **8**, Palmatolepis tenuipunctata; 08P-34 013. **9**, Palmatolepis regularis; 08P-35 015. **10**, Palmatolepis regularis; 08P-36 022. **11**, Icriodus alternatus alternatus; 08P-36 020. **12**, Gnathodus* cf. *bilineatus bilineatus; 08P-38 024.*

Sections	Las Mimbres			La Solana			
samples	06A-60	07P-260	07P-262	08P-34	08P-35	08P-36	08P-38
dissolved amount (kg)	1.375	1.325	0.900	0.200	0.400	0.550	0.600
Criteriognathus steinhornensis	X		X				
Gnathodus cf. *bilineatus bilineatus*							X
Icriodus alternatus alternatus						X	
Palmatolepis crepida				X			
Palmatolepis glabra glabra					X		
Palmatolepis minuta minuta				X	X	X	
Palmatolepis quadrantinodosalobata				X			
Palmatolepis cf. Regularis					X		
Palmatolepis tenuipunctata				X			
Pseudooneotodus beckmanni	X						
Fragments/ramiforms	17	2	19	24	34	82	-

Table II.2. Conodont elements abundance and distribution in the studied sections in the Cogollos Vega Zone.

II.3.1. Previous studies on the Western Malaguide

Stratigraphic contributions to the Malaguide Paleozoic in the Western Betic Cordillera are mainly included in regional geological studies. Among them, in addition to the seminal contributions of Blumenthal (1927), who defined this complex as Betic of Málaga, and to the studies included in the official geological maps of the region (Plan Magna maps of the IGME), the most important ones are included in several doctoral theses, mainly by german researchers. Among them that (unpublished) of Michelau (1942), those made in the sixties by students of the Bonn school, which also included detailed geological maps of this complex (Kockel, 1963; Dürr, 1967; Mollat, 1968;

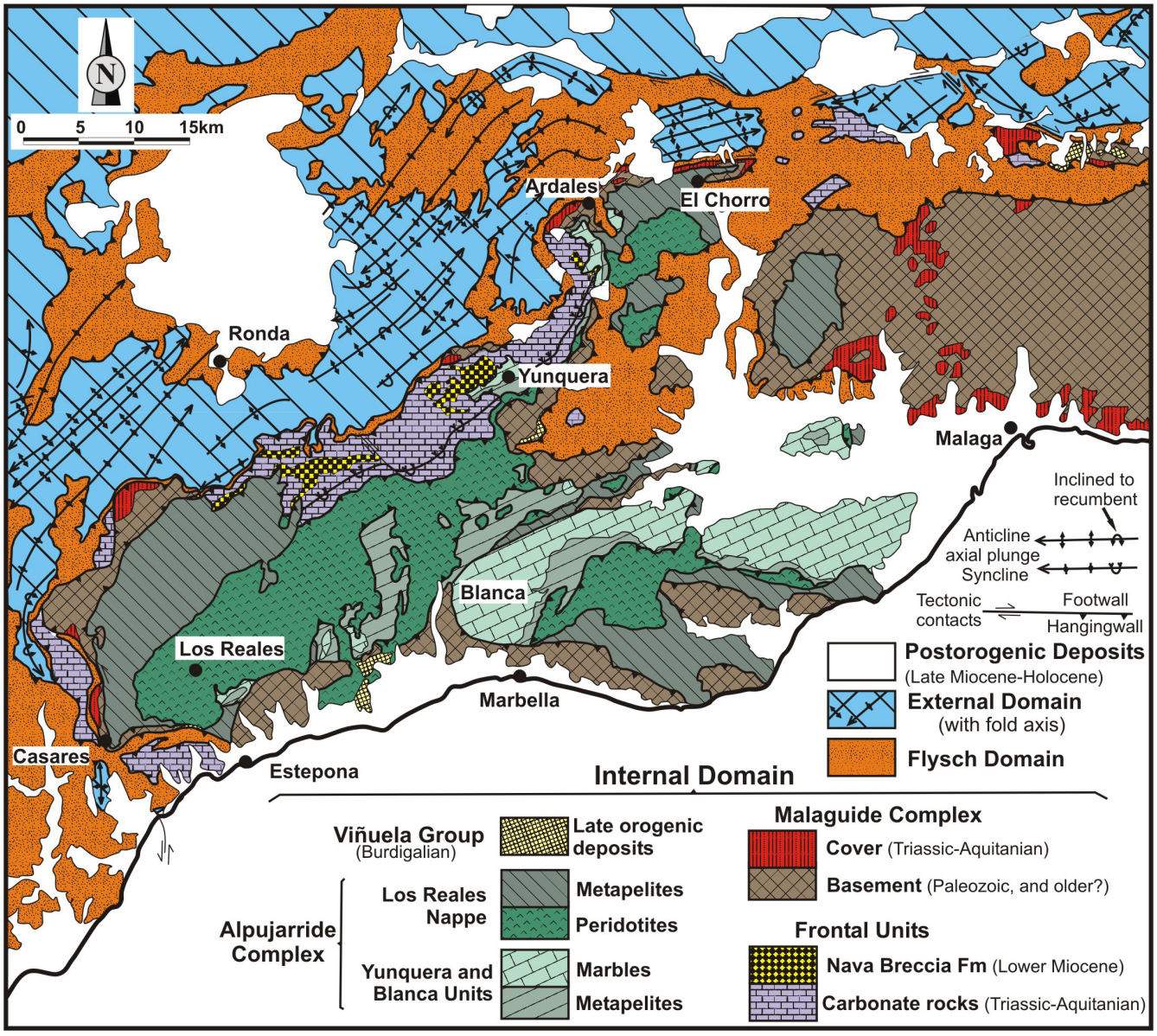


Figure II.3.1. The Malaguide Complex in the Western Betic Cordillera (from Martín-Algarra et al., 2009b)

Buntfuss, 1970) and which were also preceded by several more synthetic papers (Hoeppener *et al*. 1963, 1964; Dürr *et al*., 1960-62, 1962) and more recently, Herbig (1984) and related papers in addition to Buchroithner *et al*. (1980): Herbig, (1983, 1985, 1986); Herbig and Mamet (1983); Herbig and Stattegger (1989). Other very important contributions to the Malaguide Paleozoic stratigraphy are those of Kockel (1958, 1959), Kockel and Stoppel (1962) and Mon (1971), in addition to the thesis of Martín-Algarra (1987), who synthesized the presently accepted stratigraphic framework for the Malaguide Paleozoic, and that of Rodríguez-Cañero (1993a), devoted to the biostratigraphic study by means of conodonts of the Malaguide Complex in the province of Málaga, as well as later

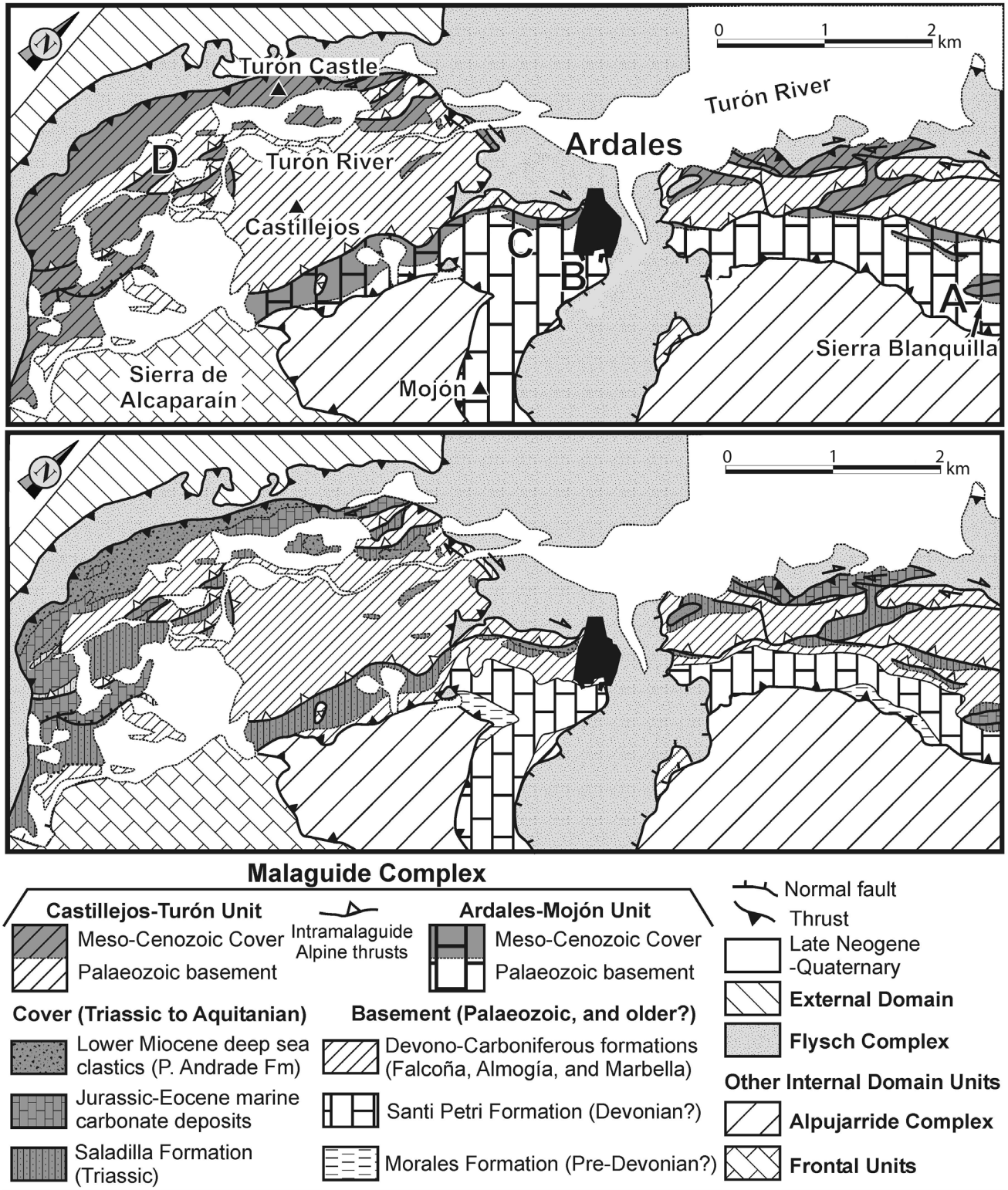


Figure II.3.2. Tectonic map and simplified geological map of the Alpine units of the Malaguide Complex at Ardales. A, B, C, and D indicate the locations of the synthetic columns reproduced in figure II.3.3 (from Martín-Algarra et al., 2009a)

papers by these and other authors (Rodríguez-Cañero, 1993b, 1995; Rodríguez-Cañero *et al*., 1997; O'Dogherty *et al.*, 2000). Most of these studies were made in the Montes de Málaga area, although an important part of them was devoted to the study of the Malaguide Paleozoic successions near Ardales, which has been the area studied in the Western Betic Cordillera for this dissertation. As one of the contributors to the research group leadered by Dr. Martín-Algarra and Dr. Rodríguez-Cañero, I have participated in some of the most recent findings on the subject in this area (Martín-Algarra *et al.*, 2009a, 2009b, Rodríguez-Cañero *et al.*, 2010, see appendix I), The main contributions of these previous studies are summarized as follows.

Blumenthal (1930, 1949) studied the Malaguide Complex in the coast to the west of Málaga. He correlated the Paleozoic successions in the Montes de Málaga with those of the Malaguide Complex in other areas of the Betic Cordillera (Blumenthal, 1928) as well as with the Paleozoic of the Rif and the Kabylides in northern Africa. Blumenthal (1930, 1949) subdivided this Paleozoic succession in several lithologic intervals, from top to bottom:

1. Polygenic conglomerates with slate and greywacke intercalations. This conglomerate is mainly composed of quartzite pebbles and limestone pebbles and boulders that include diverse marine facies and are rich in fossils, including Middle and Upper Devonian tetracorals. This interval undoubtedly corresponds to the Marbella Fm., and the corals were later dated as Carboniferous.

2. Greywackes: green-olive greywackes with plant remains (Calamites) and slate intercalations. Blumenthal (1927) ascribed this interval to the Variscan Culm facies, *i.e.*, the Almogía Fm.

3. *Calizas alabeadas*: dark bluish, massive, azoic limestones with slate and dark green greywacke intercalations. These limestones are mainly characterised by warped folds or *alabeos*. Blumenthal (1930, 1949) assigned a Silurian-Upper Devonian age to this interval based on orthoceras and dacryoconarids that he found in limestone levels supposedly equivalent to the *Calizas alabeadas*.

4. Clayey slates, mostly black and bluish. This interval probably contains slates of the Morales Fm. and other thin levels included in the Santi Petri and the Almogía Fm.

Since the contributions of Michelau (1942) and Mon (1969, 1971) these intervals are known as Marbella Fm., Almogía Fm., Santi Petri Fm. and Morales Fm. (Martín-Algarra, 1987).

Martín-Algarra (1987) mapped the Malaguide Complex terrains outcropping close to or near to the Internal-External Domain boundary in the Ronda-Ardales-Chorro and Casarabonela-Yunquera areas. He synthesized the general stratigraphy of the Malaguide Paleozoic (see also Martín-Algarra *et al*., 2004b) and recognised that in these areas the Malaguide Complex make part of three main Alpine units: lowest epimetamorphic, lower and upper Malaguide units. The Epimetamorphic Malaguide Unit is slightly touched by a very low-grade Alpine metamorphic overprint and is tectonically laminated in many areas, but present in the Convento de las Nieves Malaguide klippe (near El Burgo, 20 km to the southwest of Ardales). The Lower Malaguide Unit is regionally the most widespread, thickest, and tectonically lowest when the Epimetamorphic Malaguide Unit is absent. As mentioned in the Vélez Rubio-Lorca Corridor it is characterized by the predominance of the Morales and Santi Petri fms. Finally, he pointed out that the Upper Malaguide Unit do not include the later two formations but several Silurian-Devonian conodont-bearing horizons, as well as thick post-Paleozoic deposits.

II.3.2. Previous biostratigraphic studies

Several authors have dated different Paleozoic outcrops in the western Malaguide Complex. Blumenthal (1930) found fossils in this complex for the first time. In the Ardales area he described dacryoconarids and Upper Silurian orthoceras in limestones, which occur in a stratigraphic position he considered to be equivalent to that of the Calizas alabeadas. To the south of the Sierra de Mijas, Blumenthal (1930, 1949) found Upper Devonian trilobites, Middle-Upper Devonian tetracorals and dacryoconarids, among other fossils without biostratigraphic implications.

Kockel (1959) was the first author who found

conodonts in this complex. His work allowed to identify Silurian to Middle Devonian rocks in the Ardales and Casarabonela areas. He concluded that the limestone lenses bearing conodonts he studied were concordantly intercalated within the *Calizas alabeadas* of the Santi Petri Fm.

Later, Kockel and Stoppel (1962) studied limestone beds in the Almogía area searching for conodonts (Almogia Kalk). In this case, the limestone intercalations were not related to the Calizas alabeadas. They confirmed the presence of several Lower, Middle and Upper Devonian rocks. Some of these samples came from limestone pebbles or cobbles of the Marbella Fm. These authors also recognized other stratigraphic intervals. Few tens of metres below the *Calizas alabeadas*, they identified a conglomeratic intercalation followed by fine-grained, varicoloured and probably volcanoclastic pelites and siliceous beds, which they correlated with lithologically equivalent beds bearing Silurian graptolites found in the Moroccan Rif. On top of the *Calizas alabeadas* they observed a pelitic interval associated with greywackes that they named *Serie Rana*. This interval gradually evolves upwards to a regionally well-defined intercalation of lydites.

Buchroithner *et al*. (1980) studied the microfacies and fauna associations of the limestone pebbles within the Retamares member (Almogía Fm.) and the Marbella Fm. in the Ardales area, in the Montes de Málaga and in the Vélez Rubio-Lorca Corridor. According to the conodont fauna found, the age of the pebbles of the Retamares member ranges from Givetian (Middle Devonian) to Visean (Lower Carboniferous).

Herbig (1983) defined the Falcoña Fm. in the Montes de Málaga to the northwest of Málaga area as a marker horizon characterised by a lower radiolaritic member and, locally, an upper limestone member. He dated the limestone member as Upper Tournaisian by means of conodonts. Later, Rodríguez-Cañero and Guerra-Merchán (1996) demonstrated a Visean age for this upper limestone member.

Herbig (1984, 1985) and Herbig and Mamet (1983) studied in detail the lithofacies, the microfacies and the fauna associations in the boulders, cobbles and pebbles of both the Retamares member and the Marbella Fm. in several outcrops of the Malaguide Complex in the Costa del Sol and the Cogollos Vega Zone. The conodonts found in the limestone pebbles of the Retamares member point out the presence of middle Givetian (Middle Devonian) to lowermost Serpukhovian (Lower Carboniferous) pelagic and deep-water limestones, including a hiatus in the Upper Devonian (F/F boundary) that could be related to the Kellwasser event (Herbig, 1984, p.174). In the Marbella Fm., he described a large number of different shallow-water facies. The conodont fauna found indicates an upper Visean to lower Serpukhovian age for the limestone pebbles and boulders in the Marbella Fm.

In relation to conodont studies, the most systematic work has been made by Rodríguez-Cañero (1993a, 1993b, 1995; Rodríguez-Cañero and Guerra-Merchán, 1996; Rodríguez-Cañero *et al*., 1997). She studied, systematically searching for conodonts, limestone samples from the Paleozoic of the western Malaguide Complex. She described Silurian, Devonian and Lower Carboniferous conodonts from several outcrops in the Ardales area, in the Montes de Málaga and in the Costa del Sol. She also recognized the existence of the Kellwasser events in Almogía and accurately dated for the first time the Falcoña Fm. (Rodríguez-Cañero and Guerra-Merchán, 1996; O'Dogherty *et al*., 2000). She stated that there is not conodont-bearing limestone horizons in the *Calizas alabeadas*.

II.3.3. The Malaguide Complex at Ardales

For the purposes of the present study, the selected outcrops in the western Malaguide Paleozoic come from the Ardales region. This area is located along the main tectonic contact between the Internal and External domains in the western Betic Cordillera (figs. II.3.1 and II.3.2). Most of the obtained information has already been published in two papers and several meeting abstracts (see appendix I). One of them deals with the Variscan tectonics of the Malaguide Complex according to the obtained stratigraphic information at Ardales (Martín-Algarra *et al*., 2009a). Another contribution was devoted to the study of the first Hirnantian conodont association found in the Betic Cordillera, and to discuss its palaeogeographic implications (Rodríguez-Cañero *et al*., 2010) and it was later integrated in a third study on the Ordovician of the Iberian Peninsula (Sarmiento *et al*., 2011).

This chapter summarizes the geology of the Malaguide Paleozoic around Ardales according to these recent contributions, although I will add some complementary information, as well as some preliminary results of my research, which is still in progress and it would certainly need additional field and laboratory work. In addition to the data already published and summarized below, five new outcrops have been studied and sampled: two of them are located very close to Ardales, and make part of the Lower Malaguide Alpine Unit in the area. The other three belong to the Upper Malaguide Alpine Unit and are located to the east of the Castillejo hill and close to the Río Turón.

Field data allow to recognize, in the Ardales region, two Malaguide Alpine tectonic units, both well exposed and consisting of both a Paleozoic basement and a Mesozoic-Cenozoic sedimentary cover (Fig. II.3.2). Actually, the Alpine age of the stacking of these two Malaguide units is demonstrated by the existence of a NE-SW-striking belt of Meso-Cenozoic rocks (Fig. II.3.2). It is made of mainly Triassic continental redbeds, structurally interposed between the Paleozoic rocks of the two Alpine units. In the field, in particular, the continental Triassic rocks and Mesozoic limestones stand out between gray to dark brown pelites and sandstones for their red color and lithology, respectively. These Alpine units are hereafter named Lower Malaguide Unit (in the footwall, named Ardales-Mojón Unit by Martín-Algarra *et al*. 2009 a and b) and Upper Malaguide Unit (in the hanging wall, named Castillejos-Turón Unit by the same authors).

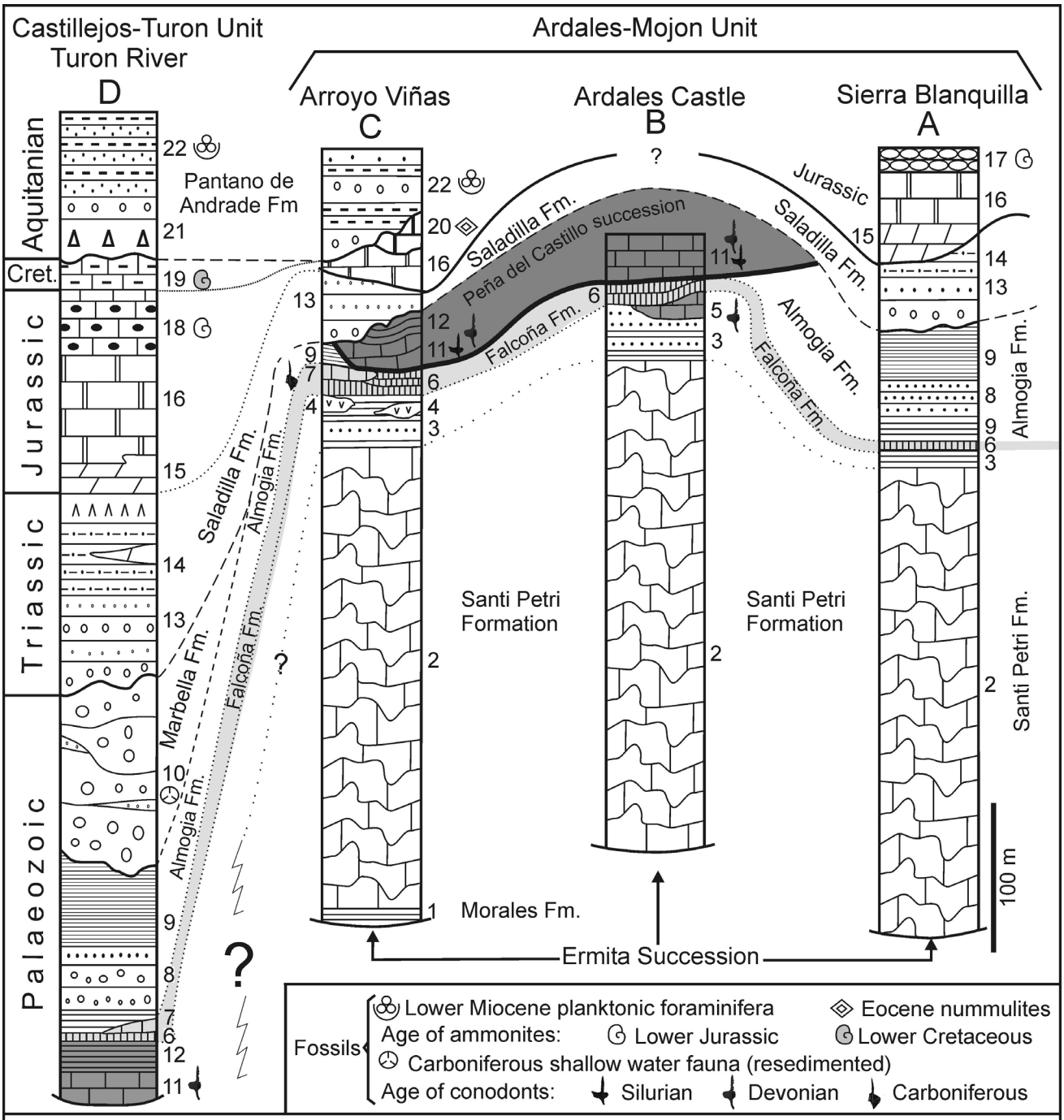
II.3.3.1. Lower Malaguide Unit

The Alpine Ardales-Mojón Unit is equivalent to the Lower Malaguide Unit defined in the Vélez Rubio-Lorca Corridor (chapter II.1) and in the Cogollos Vega Zone (chapter II.2) in this dissertation. This Alpine unit includes two superposed pre-Alpine subunits (Martín-Algarra *et al*., 2009a and b) characterized by different Paleozoic successions (both included in the Piar Group), well exposed around the Ardales Castle site (cols. B and C in Fig. II.3.3). In the following, I will refer to Ermita (lower) and Peña del Castillo (upper) successions. Post-Paleozoic sediments of the Lower Malaguide Unit also crop out near Ardales, unconformably overlying both Paleozoic successions.

Variscan Ermita Subunit

The stratigraphic succession of the Variscan Ermita Subunit crops out widely a few hundreds of metres to the northeast and southwest of Ardales (A and C in figs. II.3.2 and II.3.3) and just south of the Ardales Castle Rock (Peña del Castillo; B in figs. II.3.2 and II.3.3), around the Ermita del Calvario. One section along the Arroyo de las Viñas up to the southern foothills of the Loma de la Castaña and to the Arroyo de la Torre includes, from bottom to top, the Morales, Santi Petri, Falcoña, and Almogía fms. A small, lens-shaped outcrop of the Marbella Fm., with unclear (probably tectonic; Herbig 1983, p. 349) relations with underlying deposits, also exists 1.3 km to the south of Ardales. It is close to a fault between the Malaguide Complex (footwall) and the Campo de Gibraltar Complex (hanging wall) and it is here interpreted as a small fault-bounded drowned block belonging to the Upper Malaguide Alpine Unit.

The Morales Fm. constitutes a narrow belt, a few metres thick, of probably pre-Devonian dark gray fine-grained metapelites and metapsammites. These rocks crop out in four areas respectively located 1 km to the southwest and 2 km to the northeast of Ardales (C in Fig. II.3.3), 2 km to the northeast of Ardales (Arroyo de la Zahurda), on the western side of the Cerro del Mojón, few kilometres to the south of Ardales, and also in the Arroyo de las Cañas to the southwestern of Carratraca. South of the Ermita del Calvario, along the Arroyo de la Torre watercourse in the area to the southwest of the Loma de la Castaña, the Morales Fm. includes, at its base and near to the tectonic contact with the underlying low grade metamorphic Triassic rocks of the Alpujarride Complex, some lenses of Conglomerates with stretched pebbles. Within the same cartographic belt here described but few kilometres towards the S, along the Arroyo de las Cañas (W of Carratraca), Kockel and Stoppel (1962) recognised several decametre-thick conglomerate intercalations within the pelites and psammites equivalent to those here included within the Morales Fm. Although they are commonly strongly sheared, the rocks of the Morales Fm. show a rapid gradual lithological transition to the overlying Santi Petri Fm. by an increase in the carbonate fraction.



1. Metasandstone and schist
2. Turbiditic contorted limestone
3. Fine-grained sandstone and pelite
4. Basic volcanic rock
5. Devonian Tentaculites limestone
6. Radiolarian chert
7. Visean conodont limestone
8. Conglomerate, grauwacke and pelite
9. Olive green pelite
10. Boulder conglomerate and sandstone
11. Silurian-Devonian nodular limestone and, locally, Devonian Tentaculites limestone
12. Devonian-Carboniferous pelite and sandstone
13. Triassic red quarzitic conglomerate and sandstone
14. Red pelite, dolostone and gypsum
15. Hettangian dolostone
16. Sinemurian platform limestone
17. Pliensbachian red pelagic limestone
18. Upper Pliensbachian-Toarcian to Upper Jurassic cherty limestone and marlstone
19. Lower Cretaceous marly limestone and marl
20. Eocene Nummulites limestone
21 Aquitanian carbonate breccia
22. Turbiditic sandstone, conglomerate and pelagic marl

Figure II.3.3. Synthetic stratigraphic columns of the Malaguide Complex at Ardales (locations of columns in Fig. II.3.2) (from Martín-Algarra et al., 2009a)

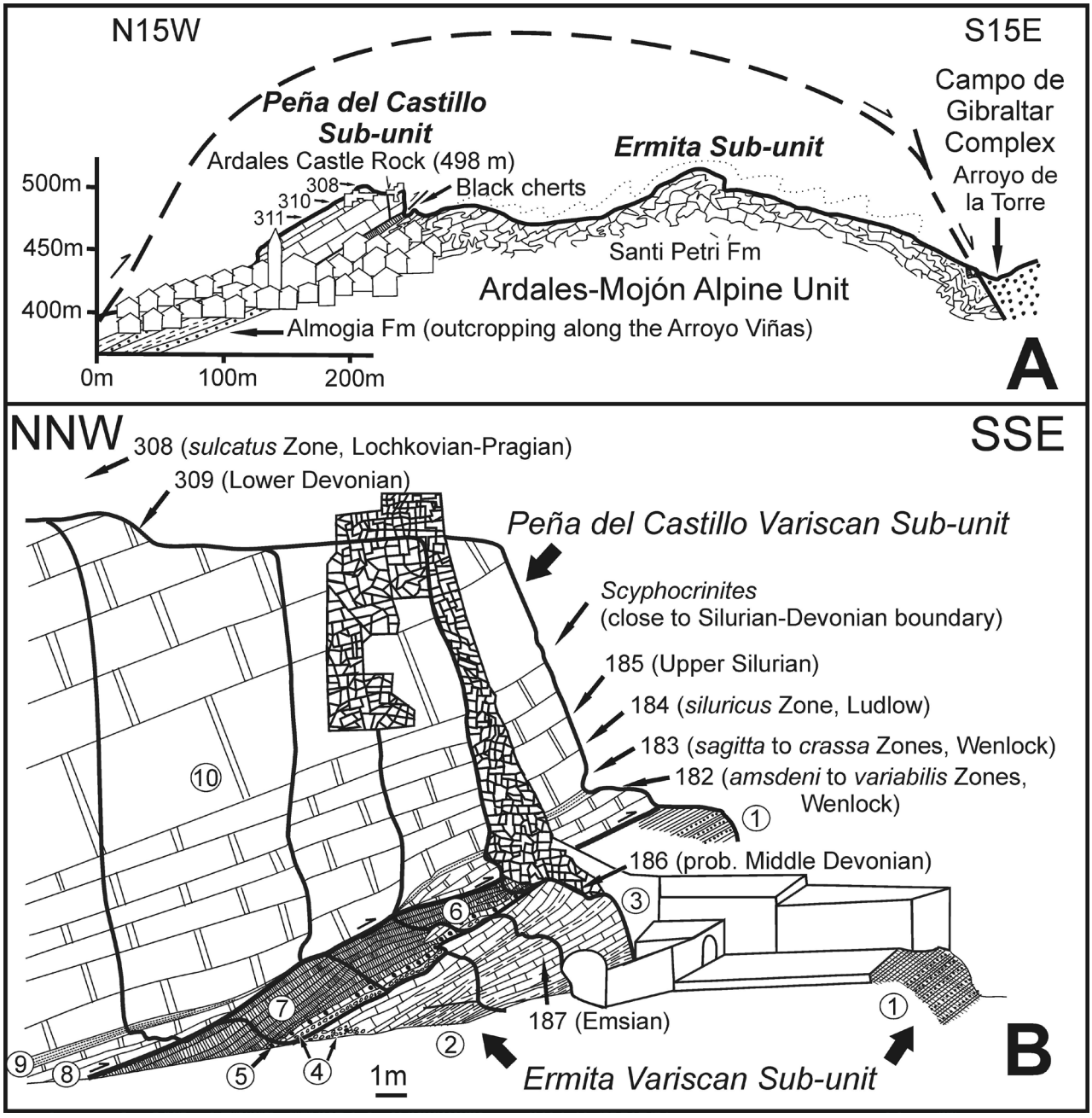


Figure II.3.4. ***A.*** *Geological section across the Ardales Castle site. This profile reproduces a natural panorama of the outcrop as it is visible from the west. The houses of the village and the outcrop of the Almogía Fm. along the Arroyo Viñas have been projected on the section.* ***B.*** *Outcrop sketch (stacked series of parallel profiles marked by thicker lines) of the southern cliff of the Ardales Castle Rock, with the location of the samples dated with conodonts (from Martín-Algarra et al., 2009a).* ***1.*** *Strongly foliated, alternating fine-grained sandstones and pelites.* ***2.*** *Olive-green pelites, including thin limestone beds and clasts and brecciated limestone lenses.* ***3.*** *Limestones with tentaculitids.* ***4.*** *Calcareous breccia.* ***5.*** *Nodules and patinas of iron oxyhydroxides included within slightly calcareous black cherts.* ***6.*** *Finely crystalline limestone block included in siliceous pelites, containing Late Ordovician conodonts.* ***7.*** *Thin-bedded, flattened, and sheared black cherts.* ***8.*** *Medium-bedded, locally strongly brecciated, Silurian nodular limestones with conodonts.* ***9.*** *Thin intercalation of black slightly siliceous shales.* ***10.*** *Nodular and massive, thick- to medium-bedded Silurian limestones.*

In the Loma de la Castaña and the Mojón belts, as well as to the NE or Ardales, the Santi Petri Fm. (2, cols. A–C, in Fig. II.3.3) is several hundreds of metres thick. It is made of intensely folded, black to gray fine-grained turbiditic limestones and calc-schists (*Calizas alabeadas* facies), associated with pelites and sandstones similar to those of the Morales Fm.

The *Calizas alabeadas* of the Santi Petri Fm. are gradually followed, in particular to the SW of Ardales, by pelites and sandstones (3 in Fig. II.3.3) with lenses of basic volcanic rocks (4 in Fig. II.3.3) and of turbiditic limestones, which have never been formally included in any specifically named lithostratigraphic unit although it should be an equivalent of the so-called *Serie Rana* by Kockel and Stoppel (1963). A Devonian-Carboniferous age may be suggested for these beds (Fig. II.3.3) because, regionally (see also below), they pass upward to the Lower Carboniferous radiolarites and limestones of the Falcoña Fm.

In the southern foothill of the Ardales Castle Rock, the top of the Ermita succession above the pelitic-psammitic interval before mentioned is formed by a laterally discontinuous lens of bluish, platy to partially nodular limestone bearing dacryoconarids, which yielded Emsian to Middle Devonian conodonts (Rodríguez-Cañero, 1993a; Martín-Algarra *et al*., 2009a; see also Fig. II.3.4). After a sharp surface, the top of the dacryoconarid limestones is covered by a laterally discontinuous breccia horizon followed by a slightly calcareous, clayey-siliceous interval rich in patinas and nodules of iron oxides. Within this horizon, which is only a few decimetres-thick, there exist one block of bioclastic sparry limestone that has provided a cool-water conodont association, Hirnantian in age (Rodríguez-Cañero *et al*., 2010, see appendix I). The most important outcrop of cherts located up to now in the Ardales area is located immediately above the former beds, at the southern foothill of the Ardales Castle Rock, constituting the top of the Ermita succession. These cherts have been attributed to the Falcoña Fm.

The Falcoña Fm. is overlain by turbiditic conglomerates, graywackes, and olive-green pelites (8 and 9 in Fig. II.3.3), which constitute the Almogía Fm. These rocks are better exposed to the northeast of Ardales (col. A in Fig. II.3.3), although they are also locally present to the southwest of the village (col. C in Fig. II.3.3; Figs. II.3.4). They are, however, much better and widely exposed in outcrops belonging to the Upper Malaguide Unit.

Variscan Peña del Castillo Subunit

Regarding to the Peña del Castillo Subunit, the upper Variscan Subunit recognised by Martín-Algarra *et al*. (2009a and b) within the Lower Malaguide Alpine Unit in the Ardales area, no new observations have been made. So the following lines summarize the stratigraphic results previously published. The Peña del Castillo succession is well exposed at the Ardales Castle Rock (Fig. II.3.4.A) and in several outcrops southwest of Ardales. It is formed by several tens of meters of well-bedded, NNW-dipping, reddish to grayish, thick- to medium-bedded nodular (griotte-like facies) and micritic limestones (11, cols. B and C, in Fig. II.3.3). This succession constitutes the best-dated Silurian–Lower Devonian succession of the Malaguide Complex (Rodríguez-Cañero, 1993a). In particular, the succession starts with medium-bedded (10–20 cm), locally strongly brecciated, brown to reddish nodular limestones (8 in Fig. II.3.4.B), followed by a thin intercalation (50–60 cm) of black slightly siliceous shales (9 in Fig. II.3.4.B). Finally, most of the Castle Rock is formed by nodular and massive, thick- to medium-bedded limestones (10 in Fig. II.3.4.B). These limestones contain orthoceratids (Blumenthal, 1930), one bed contains *Scyphocrinites* (Rodríguez-Cañero *et al*., 1997; see Fig. II.3.4.B), and they provided abundant Silurian–Lower Devonian conodonts (Rodríguez-Cañero, 1993a; Martín-Algarra *et al*., 2009a, see appendix I).

The relationships of the Silurian-Devonian succession of the Peña del Castillo with the overlying Paleozoic and younger deposits are not visible because they are hidden by the houses of the village. Nevertheless, these relationships may be observed few hundreds of metres to the west of Ardales. In this sector, Silurian and Devonian nodular limestones, sometimes bearing tentaculites and lithologically quite similar to those of the Peña del Castillo (11, col. C, in Fig. II.3.3) are locally covered by undated (Devonian-Carboniferous?) pelites and fine-grained sandstones (12 in Fig. II.3.3) and by unconformable Triassic (13 in Fig. II.3.3) and younger deposits. Silurian and Devonian limestones, well dated by conodonts (Kockel, 1959; Rodríguez-Cañero, 1993a), represent the western continuation of the Peña del Castillo succession and also tectonically overlie there the Falcoña Fm. cherts and associated rocks (6 and 7 in Fig. II.3.3) at the top of the Ermita succession.

II.3.3.2. Upper Malaguide Unit

The Castillejos-Turón Unit defined by Martín-Algarra *et al*. (2009a) is equivalent to the Upper Malaguide Unit in the Vélez Rubio-Lorca Corridor (section II.1) and in the Cogollos Vega Zone (section II.2) proposed in this dissertation. It consists of Silurian to Carboniferous rocks, represented mainly by Culm facies, post-Visean turbiditic pelites, graywackes, and conglomerates (Almogía and Marbella fms.) and by minor Lower Carboniferous cherts and limestones (Falcoña Fm.), Devonian tentaculites limestones, and Silurian nodular limestones, these latter very similar to those constituting the upper Variscan Peña del Castillo Subunit summarized in the previous section. The Meso-Cenozoic cover of this Alpine unit is thicker and stratigraphically more complete than that of the underlying Lower Malaguide Unit (Fig. II.3.3). Exactly the same is observed in the Vélez Rubio-Lorca Corridor as described in section II.1. In addition, the Jurassic to Aquitanian part of the succession is strongly tectonized and sheared along the Internal-External Domain boundary, and frequently detached from its Paleozoic to Triassic substratum.

The Morales and Santi Petri fms. are lacking in this unit, as they are exclusively present in the Lower Malaguide Alpine unit (figs. II.3.2 and II.3.3) like in the Vélez Rubio-Lorca Corridor (section II.1). The Paleozoic rocks of the Upper Malaguide Unit are represented by small outcrops of pelites and conodont-bearing Silurian-Devonian nodular to very-thin-bedded limestones (Kockel 1959). The facies of these limestones (11, col. D, in Fig. II.3.3) are quite similar to those of the Ardales Castle Rock. In the area to the NE of Castillejos, these rocks are followed by pelites and fine-grained sandstones (12 in Fig. II.3.3), including thin beds of black cherts and limestones that are similar to rocks included within the Carboniferous Falcoña Fm. (6 and 7 in Fig. II.3.3), although they have not been dated, and by the post-Visean Almogia Fm. (8 and 9 in Fig. II.3.3). The latter is especially well exposed, with its typical Culm facies of turbiditic pelites, graywackes, and conglomerates, in the southern and eastern foothills of the Castillejos hill, a few kilometres to the W of Ardales.

The top of the Paleozoic succession of the Upper Malaguide Unit is represented by the Upper Carboniferous Marbella Fm. (10 in Fig. II.3.3), which deeply erodes the underlying deposits. It is mainly formed by matrix-supported, poorly sorted, coarse and massive conglomerates (debris flow deposits), including cobbles and boulders eroded from rock successions similar to the underlying Paleozoic. Clast-supported conglomerates and finer grained facies, such as coarse-grained gravelly sandstones, are also present but only subordinately. A distinctive feature of this formation is the common occurrence of big clasts (cobbles and boulders) of limestones containing Carboniferous (Visean-Bashkirian) shallow-water fossils (Buchroithner *et al*., 1980; Herbig, 1984) and of felsic plutonic and metamorphic rocks of unknown provenance. The latter are especially well exposed to the west of Ardales along the Río Turón, but an important outcrop also exists to the south of Ardales. The Marbella Fm. outcrop located south of Ardales, at the confluence of the Arroyo de la Torre with the Arroyo de las Moratillas, is tectonically sandwiched in a fault zone between the backthrusted Campo de Gibraltar Complex and older deposits of the Lower Malaguide Unit (Variscan Ermita Subunit), and it is better interpreted as a tectonic slice belonging to the overlying Alpine Upper Malaguide Unit.

II.3.4. Studied sections

In addition to the stratigraphic and biostratigraphic data already published and summarized above, two new Paleozoic sections have been studied and sampled within the Variscan Ermita Subunit of the Lower Malaguide Alpine Unit, mainly focusing on Carboniferous rocks: one was made to the northwest of the Ermita del Calvario along the Arroyo Viñas (or Arroyo de los Ángeles), other to the south and east of it, in the southern foothills of the Loma de la Castaña and along the Arroyo de la Torre. Three Paleozoic outcrops have been studied and sampled in the Upper Malaguide Unit, although only preliminary results would be presented here, as their study is still in progress. They are located to the east of the Castillejo hill and close to the Río Turón.

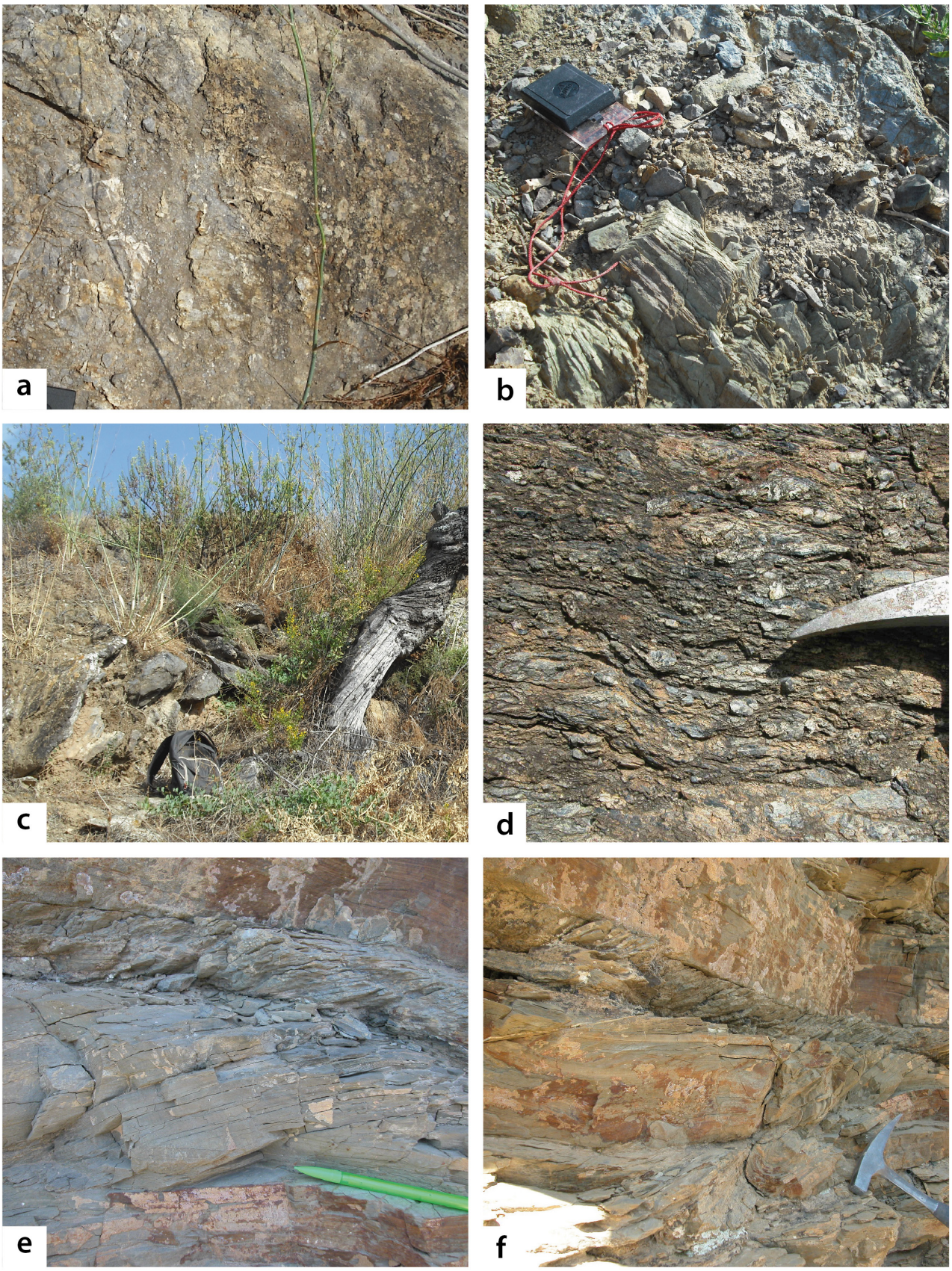


*Figure II.3.5. Arroyo Viñas (a-c) and Loma de la Castaña (d-f) areas. **a**, Calizas alabeadas in the Arroyo Viñas; **b**, Lydites and greenish pelites in the Arroyo Viñas; **c**, Platy limestones similar to those that have released Lower Carboniferous conodonts (07P-19 and 07P-20); **d**, Streched pebbles of the Morales Fm., just to the SE of the Loma de la Castaña; **e** and **f**. Strongly foliated Calizas alabeadas of the Santi Petri Fm. in the surroundings of the Ermita del Calvario, to the south of Ardales.*

II.3.4.1. Lower Malaguide Unit

As mentioned above, in the Variscan Ermita Subunit to the southwest of Ardales the top of the Santi Petri Fm. is defined by a mappable horizon, a few tens of metres thick, made of slates and greenish sandstone beds, which intercalates laterally discontinuous lenses of greenish volcanic rocks (Fig. II.3.5.a and b). In the eastern side of the Arroyo Viñas (Fig. II.3.6, see location of the section in Fig. II.3.3), within these beds, a laterally discontinuous small outcrop of typical black cherts belonging to the lower radiolaritic member of the Falcoña Fm. crops out dipping to the north. Overlying these beds, and in stratigraphic contact with them, one lens of platy bluish limestones appear (Fig. II.3.5.c). The samples taken from this lens, which is less than 1 m thick, have not yielded any conodont remain; however, two samples taken from a glided block to the bottom of the gully, a few metres apart from the above mentioned outcrop, have yielded conodonts which indicate a Lower Carboniferous age (see section II.3.5 for biostratigraphic details). This supports the Carboniferous age proposed by Martín-Algarra *et al*. (2009a) for the strongly deformed radiolarian cherts appearing in the same cartographic belt and stratigraphic position (with respect to the underlying Santi Petri Fm.) at the base of the Ardales Castle Rock, delineating the top of the Variscan Ermita Unit. These preliminary biostratigraphic results were briefly mentioned by Martín-Algarra *et al*. (2009a), and have been already published as a meeting abstract (Navas-Parejo *et al*., 2012a, see appendix I).

Above the Falcoña Fm. nearby to Ardales, the outcrops of Almogía Fm. are small and poorly exposed, and it is sometimes unclear if they belong to the Lower or to the Upper Alpine Malaguide units. The turbiditic conglomerates, graywackes, and olive-green pelites (8 and 9 in Fig. II.3.3) of the Almogía Fm. are better exposed to the northeast of Ardales (col. A in Fig. II.3.3), and also (but strongly tectonized) to the south of the village, along the Arroyo de la Torre.

The second outcrop studied in the Ermita Subunit is located to the south of the Ermita del Calvario, between the Loma de la Castaña and the Arroyo de la Torre. There, the stratigraphic succession of the Variscan Ermita Subunit is difficult to establish in detail due to strong tectonics (probably both Alpine and pre-Alpine). Along the track from the Ermita to the Cerro del Olivo, the *Calizas alabeadas* (Fig. II.3.5.e and f) form the reverse limb of a NE-SW trending narrow anticline and, a few tens of metres to the south, a different succession, probably reversed but yet dipping to the north, appears geometrically below typical *Calizas alabeadas* facies of the Santi Petri Fm. in the southern side of the Loma de la Castaña: it is formed by a few metres of strongly folded but well bedded bluish limestones with chert beds and nodules, which appear intercalated between greywackes and slates (Fig. II.3.7.b and c). These beds strongly remind some facies of the Visean limestones corresponding to the upper member of the Falcoña Fm., as they appear in some areas of the Montes de Málaga (e.g. Pechuga section: see O'Dogherty *et al*., 2000). Unfortunately, the typical cherts of the Falcoña Fm. have not been observed here, and none of the three samples taken from these limestones have provided any biostratigraphic information. Consequently, it cannot be excluded that they could constitute beds stratigraphically underlying the Santi Petri Fm., as badly exposed outcrops of the Morales Fm. are not far from here (Fig. II.3.5.d). To the NE of this outcrop, at approximately 300 m and along the Arroyo de la Torre watercourse, white quartzites strongly deformed crop out (Fig. II.3.7.a). They most likely correspond to the radiolaritic member of the Falcoña Fm. A limestone level attached to such quartzites was sampled although did not release any conodont remain.

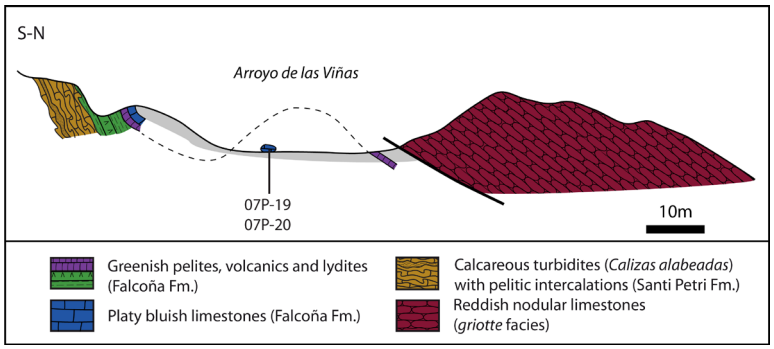


Figure II.3.6. Arroyo Viñas section.

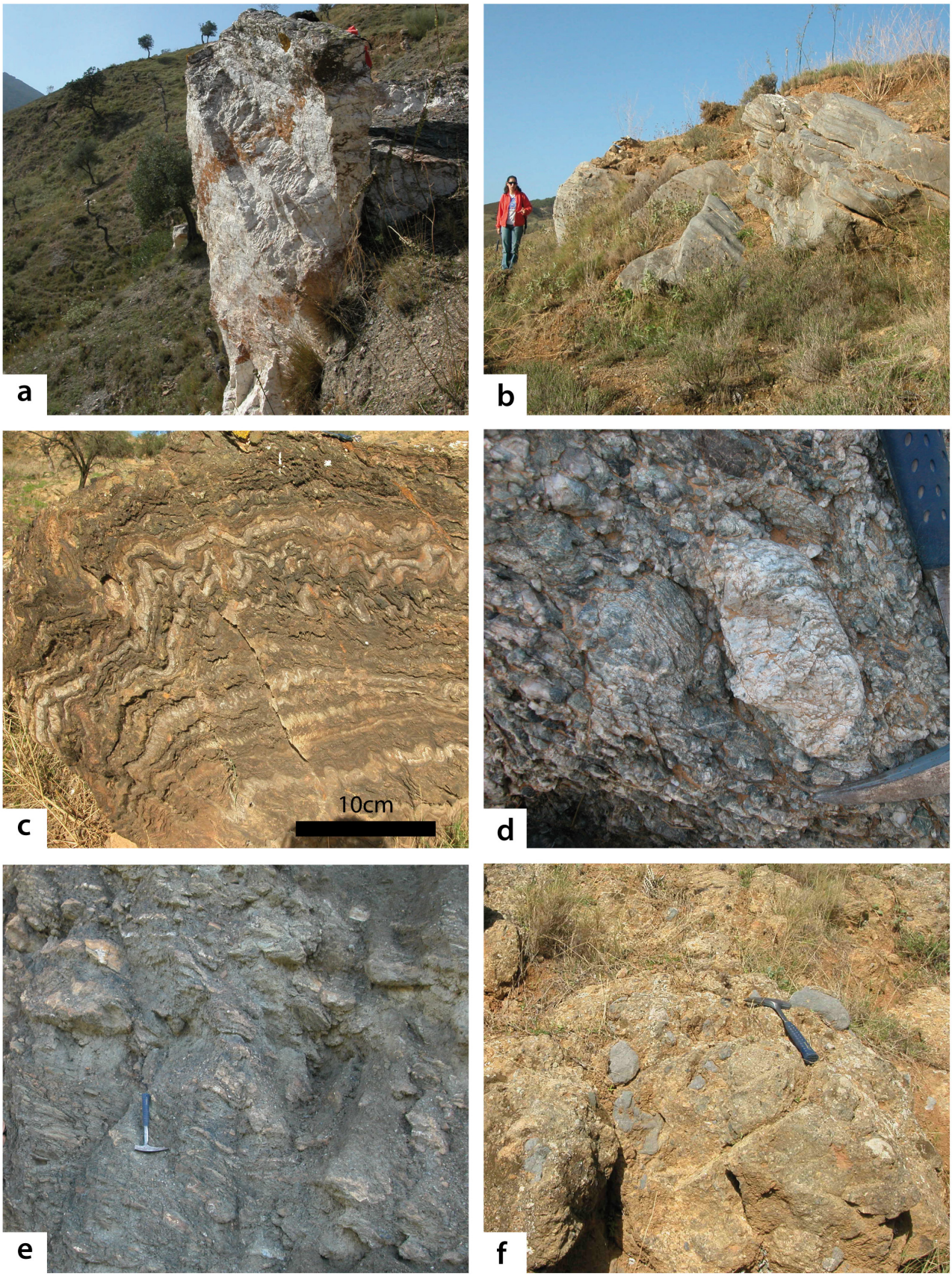


Figure II.3.7. Arroyo de la Torre and Loma de la Castaña areas. ***a****, White quartzites in the Arroyo de la Torre;* ***b****, Massive bluish limestones with chert levels and nodules in the Loma de la Castaña;* ***c****, Strongly deformed limestones with chert levels;* ***d****, Granite clast in the Retamares conglomerate of the Almogía Fm.;* ***e****. Slates with thin sandstone intercalations of the Almogía Fm.;* ***f****. Conglomerate of the Marbella Fm. in the Arroyo Moratillas.*

II.3.4.2. Upper Malaguide Unit

About 1 km to the south of the Turón Castle, between the Arroyo de Cerro Romero and the Arroyo del Tranco, a few hundreds metres of Paleozoic rocks of the Upper Malaguide Unit are tectonically sandwiched between Triassic deposits of the Saladilla Fm. and Jurassic brecciated limestones. Just to the south of the Loma del Castillo, on the left side of the Turón River, in the surroundings of a new farmhouse, massive nodular limestones occur (Fig. II.3.9.a and b). These limestones have released some conodonts whose age ranges from the Middle Ordovician to the Lower Devonian (see II.3.4. for biostratigraphic details). However, their facies resembles that of other nodular limestones more accurately dated as Silurian to lowermost Devonian in the Ardales Castle succession (Rodríguez-Cañero, 1993a; Rodríguez-Cañero *et al.*, 1997; Martín-Algarra *et al*., 2009a).

To the north, along a forestal track parallel to the Río Turón (Fig. II.3.8), the Paleozoic succession starts with few tens of metres of strongly N-dipping platy limestones intercalated in dark slates (Fig. II.3.9.c and d). These platy limestones bear some dacryoconarids, and have released Lower Devonian conodonts (see II.3.4. for biostratigraphic details). After a few tens of metres of covered succession, a conglomerate with typical Retamares facies, including some clasts of granite and limestone, and followed by dark grey pelites and greywackes (Fig. II.3.9.e and f), can be certainly assigned to the Almogía Fm.

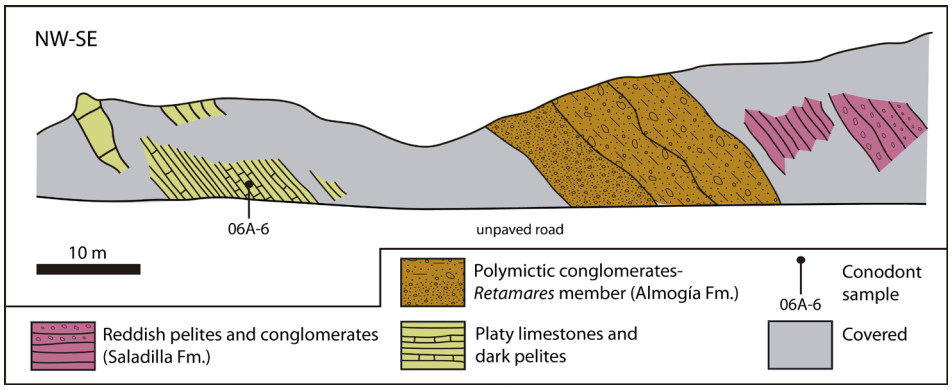


Figure II.3.8. Turon 1 section.

II.3.5. Conodont biostratigraphy

From a total of twenty five samples that were taken in the Ardales area for conodont studies, four have provided useful biostratigraphic information. Devonian (and even older) conodonts have been found in the Upper Malaguide Unit (Río Turón 1 and 2 sections). In the Lower Malaguide Unit, Lower Carboniferous conodonts have been found in this area for the first time (Arroyo Viñas section; Navas-Parejo *et al*., 2012a). Total dissolved amount of sample and detailed taxonomic results are given in table II.3.

In the Arroyo Viñas section (Fig. II.3.6), two samples from a total of three have yielded conodont remains (07P-19 and 07P-20). Although these conodonts are broken and very badly preserved, they can be attributed to Pa elements of the Lower Carboniferous genus *Gnathodus* Pander 1856, since the distinctive parapet of this genus can be clearly observed (Fig. II.3.10.1 and 2). Furthermore, the conodont element illustrated in Fig. II.3.10.3 preserves part of the outer platform, where aligned nodules parallel to the carina can be observed. Therefore, the presence of *Gnathodus bilineatus bilineatus* (Roundy, 1926) is confirmed. This indicates an age that compises the *bilineatus-bollandensis* zones (Visean-Serphkovian; Lower Carboniferous). This age is reinforced by the presence of *Lochriea commutata* (Branson and Mehl, 1941) in one of the samples (Fig. II.3.10.4).

Four samples were taken in the platy limestones described in the previous section which are part of the Upper Malaguide Unit and crop out close to the Río Turón. Only one sample (06A-6) has released *Criteriognathus steinhornensis* (Ziegler,

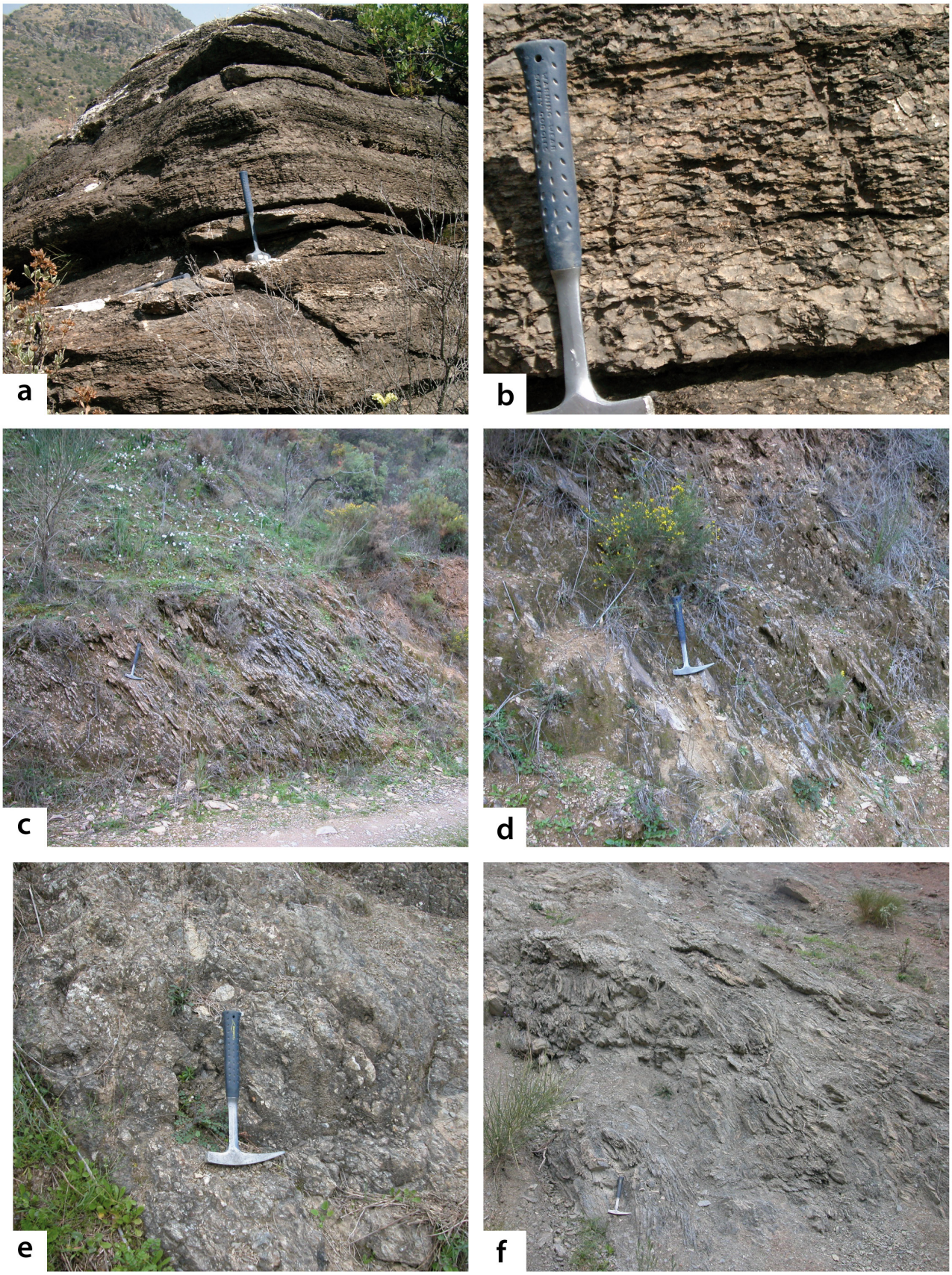


*Figure II.3.9. Río Turón sections. **a** and **b**, Massive nodular limestone which make part of the Upper Malaguide Unit; **c** and **d**, Platy limestones to the south of the Castillo Turón; **e**, Retamares conglomerate of the Almogía Fm.; **f**, Slates of the Almogía Fm.*

1956), which indicates an Emsian age (uppermost Lower Devonian). Pa elements of *Criteriognathus steinhornensis* are characterised by two clearly asymmetric lateral lobes (Fig. II.3.10.5 and 6).

In the nodular limestones cropping out close to the previous section dated as Emsian, three samples were taken. One sample (06A-54) has yielded *Pseudooneotodus beckmanni* Bishoff and Sannemann, 1958, which is characterised by coniform elements that have a curved peak with an anterior groove that looks like a hump (Fig. II.3.10.7 and 8). This species has a long-lasting stratigraphic distribution ranging from Middle Ordovician to Lower Devonian. However, similar facies have been dated as Silurian in other outcrops, as in the Peña del Castillo section or in the Arroyo Viñas (Rodríguez-Cañero, 1993a; Rodríguez-Cañero *et al*., 1997). Further sampling and conodont studies are needed in order to confirm this correlation.

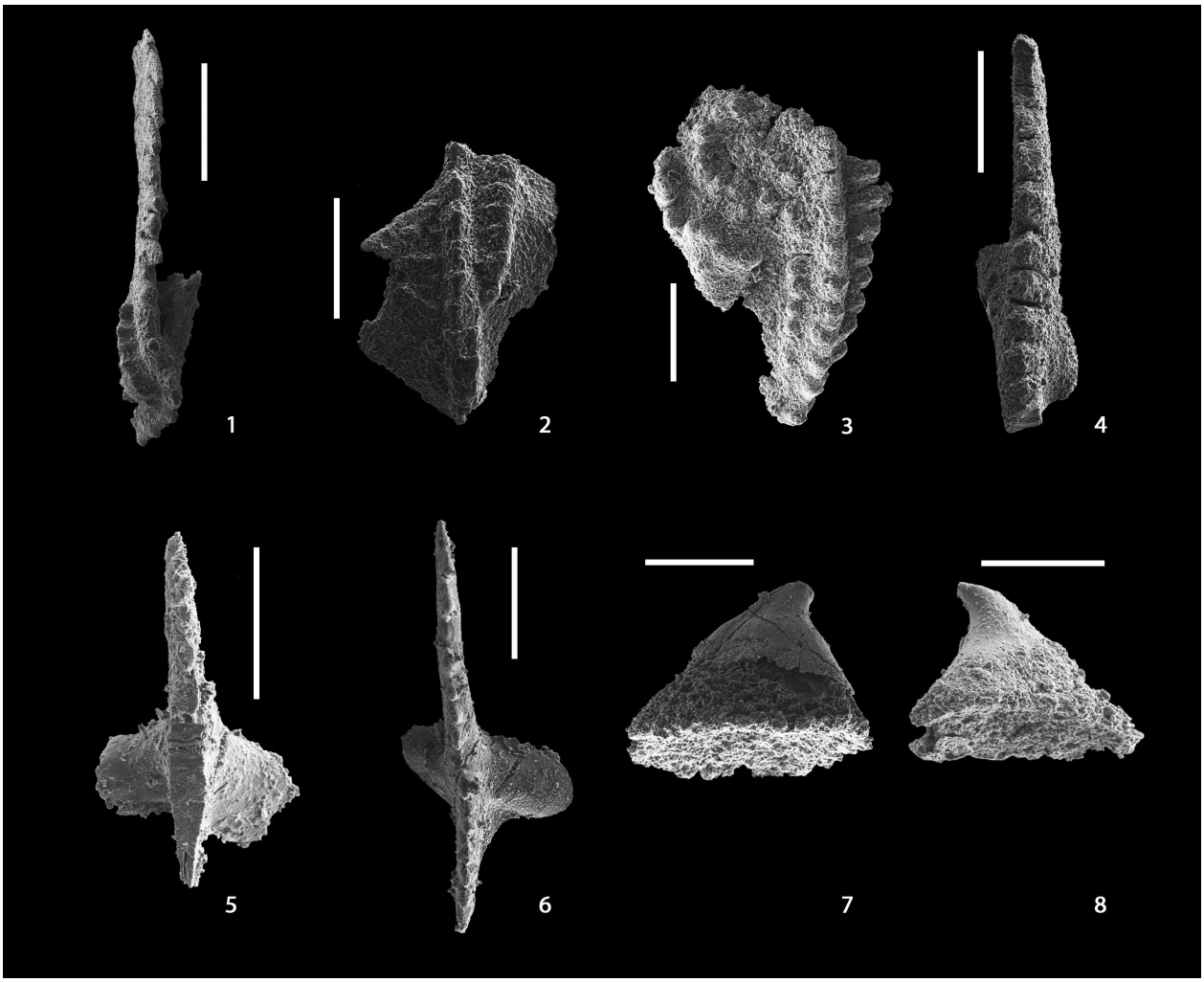


Figure II.3.10. Conodonts in the Ardales area. All are Pa elements. Scale bars are 200 µm.
***1**, Gnathodus sp.; 07P-19-1. **2**, Gnathodus* sp.*; 07P-20-3. **3**, Gnathodus bilineatus bilineatus; 07P-20-4. **4**, Lochriea commutata; 07P-20-5. **5**, Criteriognathus steinhornensis; 06A-6-2. **6**, Criteriognathus steinhornensis; 06A-6-1. **7**, Pseudooneotodus beckmanni; 06A-54-1. **8**, Pseudooneotodus beckmanni; 06A-54-2.*

Sections	Arroyo Viñas		Río Turón 1			Río Turón 2	
samples	07P-19	07P-20	06A-5	06A-6	06A-8	06A-53	06A-54
dissolved amount (kg)	1.375	0.900	0.400	0.200	0.250	0.300	0.400
Criteriognathus steinhornensis				X			
Gnathodus bilineatus bilineatus		X					
Lochriea commutata		X					
Pseudooneotodus beckmanni				X			X
Gnathodus sp./fragments	1	5					
Fragments/ramiforms	14	28	3	58	3	3	70

Table II.3. Conodont elements abundance and distribution in the studied sections in the Ardales area.

II.4. Synthesis of the Malaguide Paleozoic stratigraphy

II.4.1. Introduction

Traditionally the Paleozoic successions of the Malaguide Complex constitute the Piar Group, which is formed by the Morales, Santi Petri, Falcoña, Almogía and Marbella formations, as summarized in Chapter I. Nevertheless, there exists other rock successions that actually do not fit within any of the above mentioned classical formations, among them most of conodont-bearing limestone horizons found, up to now, in the Malaguide Complex. This is the case, apart from the limestones that constitute the upper member of the Falcoña Fm., of the different pelagic carbonate lenses and isolated beds between pelites, of Late Ordovician to Late Carboniferous age, which have provided conodonts and dacryoconarids, among other macro and microfossils in this study (figs. II.4.4 and II.4.5). Consequently, the stratigraphic frame of the Malaguide Paleozoic is much more complicated than that accepted today, and must be completed with new lithostratigraphic units that would have to be defined in a near future. In addition, it becomes evident from this study that the regional distribution of these stratigraphic units, either if formally defined or not, is not homogeneous in the different Malaguide tectonic units that have been identified in the three Malaguide areas studied for this dissertation: Vélez Rubio-Lorca Corridor (section II.1; Fig. II.4.1), Cogollos Vega Zone (section II.2; Fig. II.4.2), and Ardales area (section II.3; Fig. II.4.3). These three areas are close to the boundary between the Internal and the External Zones of the Betic Cordillera, which is a strongly tectonized region.

Consequently, the reconstruction of Malaguide stratigraphy in these areas, and the correlation between the different successions recognised in them is a difficult task. Nonetheless, the field and stratigraphic study previously presented reveals some common tectonic and stratigraphic features of the Malaguide Complex in these three areas, as well as some minor differences. In the next lines the available stratigraphic information will be synthesized, taking into account the tectonic subdivision of the studied Malaguide terrains, on the base of the stratigraphic correlation shown in figs. II.4.1, II.4.2 and II.4.3.

In spite of the strong tectonic disturbance of the studied Paleozoic terrains, this attempt allows to recognise some regional tendencies in lateral stratigraphic and facies changes. This helps to improve the available palaeogeographic scheme about Malaguide successions and to better interpret the main features of their geohistorical evolution. Moreover the proposed correlation scheme allows discussing the observed lateral and vertical stratigraphic and facies changes in the Malaguide Paleozoic within the paleogeographic frame of an Ordovician to Early Carboniferous continental palaeomargin that opened towards SE and that was closed in mid- to Late Carboniferous time, at the end of the Variscan Orogeny. This palaeomargin was related to the Paleotethys ocean, as it was also the case of most Paleozoic succession in more eastern Alpine realms.

II.4.2. Alpine vs. pre-Alpine Malaguide units

In the three areas studied, the Malaguide Complex is formed by several independent tectonic units. These units were tectonically individuated, mainly, during the Alpine cycle, as the sediments of the Meso-Cenozoic cover are also involved in the Alpine thrust contacts together with Paleozoic rocks of the pre-Alpine basement. Nevertheless, at least in one case (Ardales area), it is demostrated

that the Malaguide Complex also contains pre-Alpine thrust units. These pre-Alpine units were individuated during the Variscan Orogeny, which also leaved an intense tectonic imprint in Malaguide terrains, which is clearly recorded by the much stronger deformation of the Paleozoic successions when compared to that of the Meso-Cenozoic successions.

The Alpine Malaguide units in the three studied areas constitute narrow belts formed by thrust slices bounded by subvertical to N-dipping contacts whose kinematics in not yet well undestood. Their trends run subparallel to the regional structural strike, which is NNE-SSW in the Vélez Rubio Corridor and in the Cogollos Vega Zone, and NE-SW in the Ardales area. Three groups of Alpine Malaguide imbrications have been distinghished, and they have been grouped in three regionally well documented tectonic units according to: i) their geographic and tectonic location within the thrust stack, ii) the general features of their stratigraphic successions, and iii) the progressively stronger deformation and metamorphic recrystallization observed towards the tectonically lower units. These are, from botton to top: the Epimetamorphic Malaguide Unit, the Lower Malaguide Unit, and the Upper Malaguide Unit, which constitute, respectively, a southern, a central and a northern belt. The Paleozoic series of all these units is usually found in normal stratigraphic succession, although reversed successions dipping to the N (top towards the S) have also been recognized in the Upper Malaguide Unit of the Cogollos Vega Zone, probably as a result of Alpine backfolding or backthrusting.

The outcrops of the Alpine Malaguide Units in the studied areas are usually around a few hectometres wide (although locally they are up to a few kilometres wide, at most). They show moderate general stratigraphic and facies variations along strike for kilometres. Nevertheless, tectonically induced strong lateral variations in thickness, and in cartographic distribution of different formations, are observed in both the Paleozoic and the Meso-Cenozoic terrains of all these units, up to make them dissapear completely in some areas. This commonly accounts for the lowermost Epimetamorphic Malaguide Unit, but also for the other units, either Alpine or pre-Alpine. Secondary, laterally discontinuous Alpine thrusting with limited heave, throw and lateral continuity, is particularly evident in the Upper Malaguide Unit, which can be subdivided in two or more slices according to their tectonic features and details of their Meso-Cenozoic stratigraphy. The Meso-Cenozoic cover of the Upper Malaguide is usually the best developed and thickest, especially in its post-Triassic part. On the contrary, the Mesozoic cover of the Epimetamorphic and Lower Malaguide Units in all sectors is formed mainly by thick Triassic redbeds and subordinated dolostones of the Saladilla Fm., whereas post-Triassic sediments are absent or, when present (only in the Lower Malaguide Unit), they are extremely thin, condensed, and full of stratigraphic gaps.

From the mentioned above Alpine Malaguide Units, the Lower Malaguide Unit is usually the widest outcropping, and that bearing the standard Paleozoic succession formed by the Morales, Santi Petri, Falcoña and Almogía formations, but not by the Marbella Fm. This is exclusively preserved, in the studied areas, in the Upper Malaguide Unit. In addition, the succession of the Upper Malaguide Unit never contains the Morales or the Santi Petri fms., which have been replaced by the limestone horizons intercalated between pelites, and that have been well dated by conodonts. These conodont-bearing Silurian-Devonian limestones are exclusive of the Upper Malaguide Unit, with two remarkable exceptions at Ardales and at the Cortijo de los Gonzalez. The González outcrop makes part, cartographically, of the lowermost Epimetamorphic Malaguide Unit but is tectonically detached and isolated from the underlying normal stratigraphic succession of this unit. It bears Devonian and Bashkirian carbonate horizons, in addition to a Visean carbonate lens that belongs to the upper member of the Falcoña Fm. At Ardales, conodont-bearing limestones have been found within two tectonic subunits in which the Alpine Lower Malaguide Unit is subdivided.

In the Ardales area, in addition to the Alpine tectonic structure described above, the existence of pre-Alpine tectonic features in the Malaguide Complex has been recently demonstrated. There, unconformably below the Triassic succession of the Lower Malaguide Unit, two tectonically independent and stratigraphically different Paleozoic rock successions (Martín-Algarra et al. 2009a) are bounded by a pre-Alpine thrust whose kinematics indicates a top-to-the-SE sense of shear (Martín-Algarra et al. 2009b). Consequently, the Alpine structure of the Malaguide Complex overprints a pre-Alpine tectonic stack of Variscan thrust sheets. Interestingly, the Silurian-Devonian stratigraphic succession of the Variscan Upper Subunit at Ardales is quite similar to the Silurian-Devonian succession

commonly found in the Alpine Upper Malaguide Unit, either at Ardales or in the other three studied areas, whereas that of the Variscan Lower Subunit is the same as that of the Alpine Lower Malaguide Unit in all areas. This indicates that the Paleozoic stratigraphy of the Alpine Malaguide Units bears a strong tectonic signal inherited from the pre-Alpine structure, which was formed by stacked thrust units with different stratigraphy and pre-Triassic regional distribution. Consequently, although the tectonic pre-Alpine imprint is only relatively well preserved at Ardales, in the other areas it can be also inferred from the regional distribution of the Paleozoic successions of the Alpine Malaguide Units. The cartographic distribution of the stratigraphic features of the Paleozoic successions in different belts evokes that of former different sectors of the original sedimentary basin. In the studied area, and in present day coordinates, these features run along the strike of the Betic Cordillera, then roughly following a NE-SW direction.

In brief, and apart from the two above mentioned cases, the Paleozoic succession of the Upper Malaguide Unit shows appreciable differences with respect to that of the Lower and of the Epimetamorphic Malaguide units during both Silurian-Devonian and Late Carboniferous. In Early to mid-Carboniferous time, however, the stratigraphic succession of all Malaguide units is fromed by the Falcoña and Almogía formations, whose stratigraphic and facies features are quite similar in outcrops assigned to different Alpine Malaguide Units. Nonetheless, in the Upper Malaguide Unit these two formations are usually less strongly deformed and recrystallized than in the Lower and Epimetamorphic Malaguide Units.

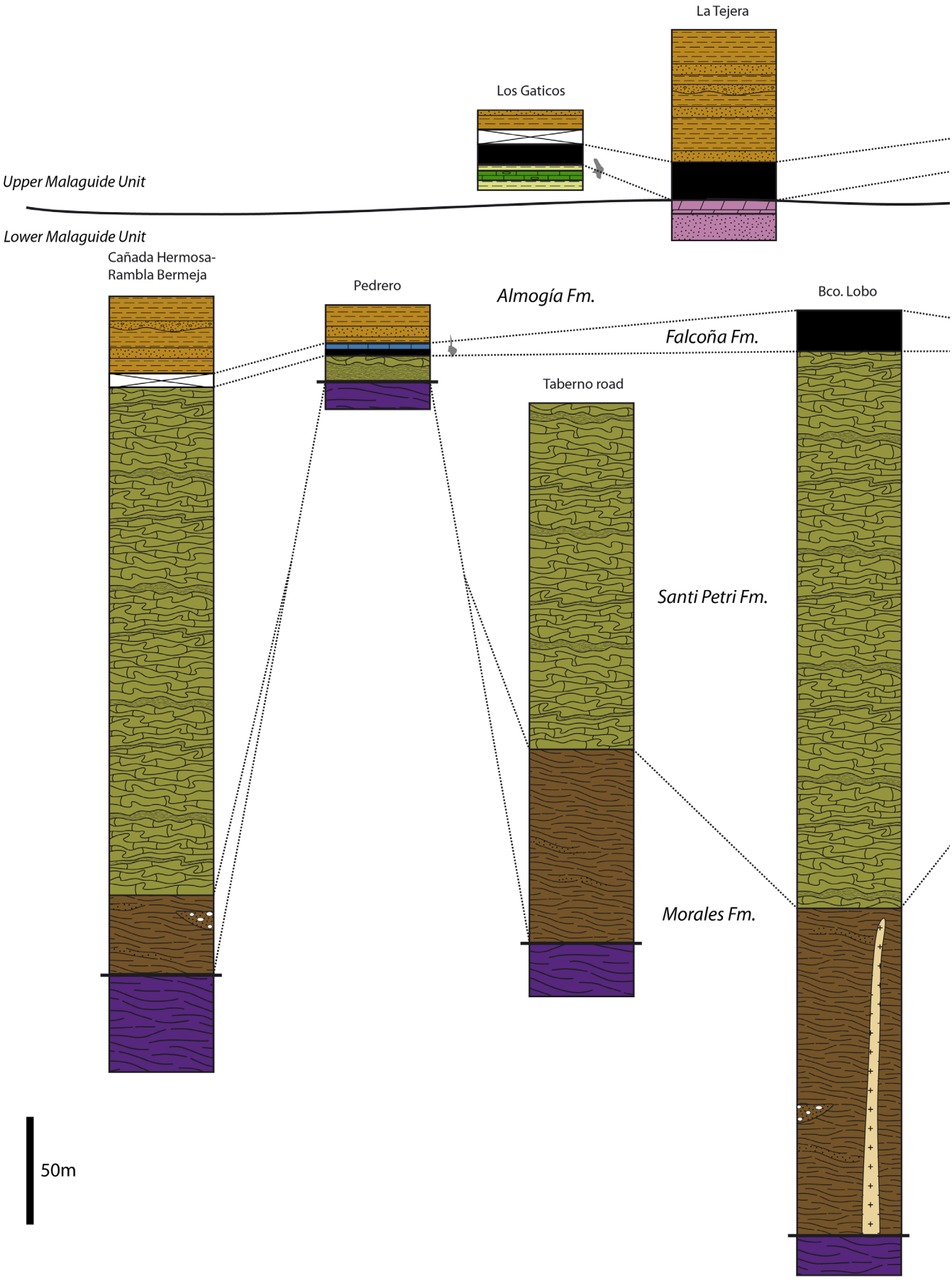
II.4.3. Epimetamorphic Malaguide Unit

The Epimetamorphic Malaguide Unit appears in the lowermost tectonic position, above typical Alpujarride Triassic terrains and below typical Paleozoic Malaguide terrains. Some of their outcrops in the studied and in other areas of the Betic Cordillera have been considered to belong to the Alpujarride Complex, or to tectonic units with intermediate stratigraphic and tectononometamorphic features between those of the Malaguide and the Alpujarride complexes (Transitional Units of Sanz de Galdeano et al., 1995, in the Cogollos Vega Zone). The Epimetamorphic Malaguide Unit has not been found in the Ardales area, probably due to tectonic elision.

The stratigraphic succession of the Epimetamorphic Malaguide Unit is mainly made of Triassic rocks (dark dolomites, silty slates, reddish-pinkish metasandstones and metaconglomerates), which are identical to equivalent rocks of the Saladilla Fm. in spite of the Alpine metamorphic recrystallization that affects them (Roep, 1972). In a lesser extent, the Epitamorphic Malaguide Unit also includes Paleozoic dark pelites and greywackes that also show indubitable Malaguide features in spite of their moderate Alpine metamorphic overprint. These Paleozoic metapelites have been found in the Cogollos Vega Zone and in the Vélez Rubio-Lorca Corridor, although stratigraphic studies about them have only been performed in the last sector (Cortijo de los González area: Fig. II.4.1).

The Epimetamorphic Malaguide Unit was first distinghished by Roep and Mac Gillavry (1962) in the Vélez Rubio-Lorca Corridor with the name of Casolidad Unit of their Betic of Malaga (later named Malaguide Complex). The Malaguide affinity of the Epimetamorphic Malaguide Unit is beyond any doubt according to the nature of its Triassic succession. In addition, ocassionally, the Paleozoic rocks of the Epimetamorphic Malaguide Unit yet preserve identifiable sedimentary features that allow their correlation with typical Malaguide formations like the Morales, Santi Petri and Almogía Fms. Diabase intrusives like those of the classical area of the Montes de Málaga are also present within Morales- and Santi Petri-like terrains of this unit, as well as a few badly preserved lyditic remnants attributed to the Falcoña Fm (i.e., González 1 section, Vélez Rubio-Lorca Corridor).

Fossiliferous beds in the Paleozoic rocks of the Epimetamorphic Malaguide Unit have been found for the first time in the surroundings of the Cortijo de los González, in the Vélez Rubio-Lorca Corridor (González 1, 2 and 3 sections, see chapter II.1.3). This represents a further support for attributing the terrains included within this unit to the Malaguide Complex, and not to the Alpujarride Complex. But, paradoxically, the González outcrop shows features


La Tejera
Los Gaticos
Upper Malaguide Unit
Lower Malaguide Unit
Cañada Hermosa-
Rambla Bermeja
Pedrero
Almogía Fm.
Bco. Lobo
Falcoña Fm.
Taberno road
Santi Petri Fm.
Morales Fm.
50m

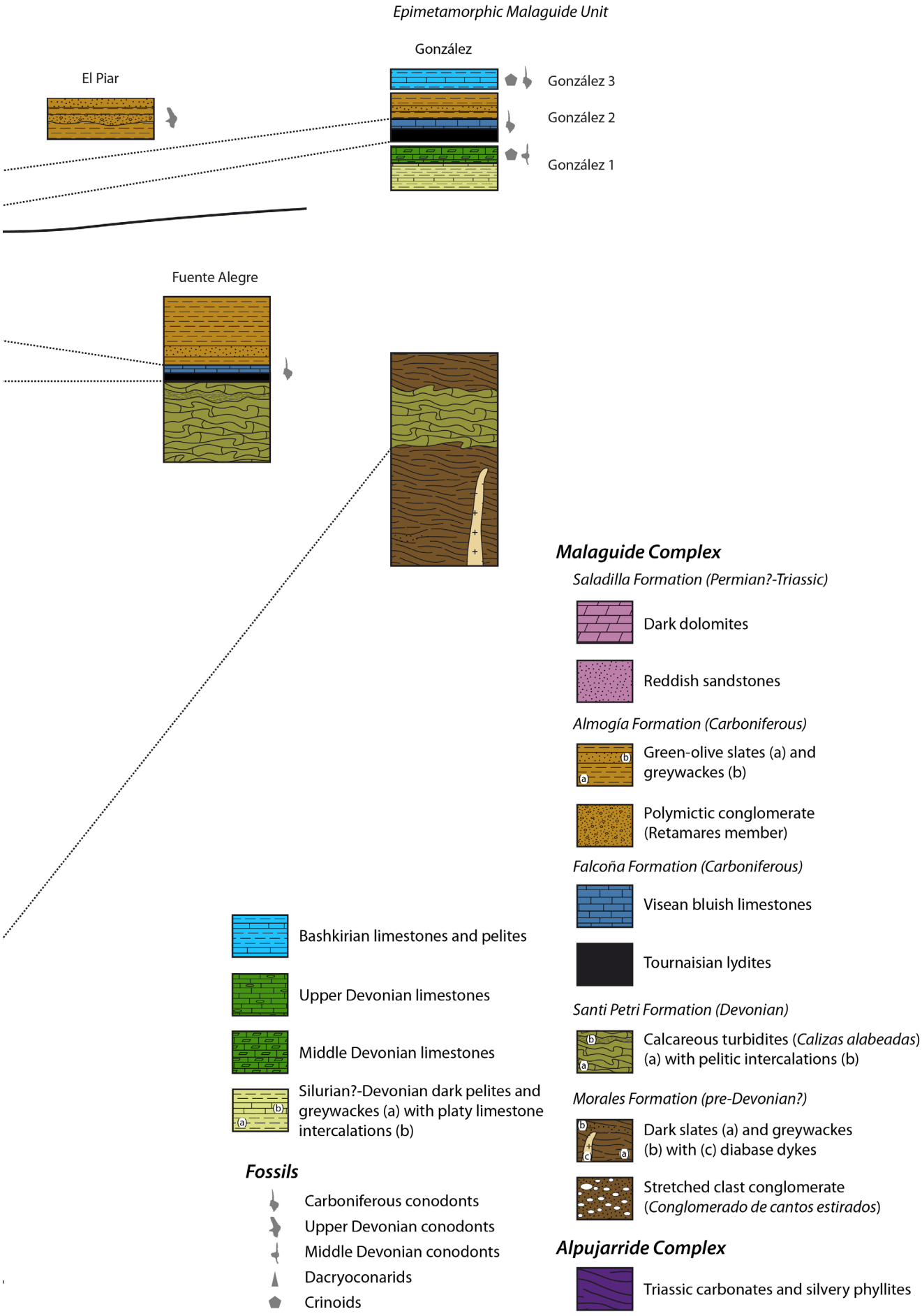


Figure II.4.1. Stratigraphic correlation chart of the Malaguide Paleozoic in the Vélez Rubio-Lorca Corridor.

that remember much more the coeval Devonian-Carboniferous beds of the Upper Malaguide Unit than the underlying Paleozoic terrains of the Epitamorphic Malaguide Unit, above which it is tectonically detached. Actually the Cortijo de González constitutes a tectonically independent and very particular succession, quite atypical when compared to other Paleozoic outcrops in the surroundings.

The Middle Devonian limestones of González 1 section occur in a stratigraphic position below Lower Carboniferous dated beds of the Falcoña Fm. Then, they occupy a stratigraphical position equivalent to that of the Santi Petri Fm. But the González Devonian beds are not stratigraphically related to the rocks, showing facies equivalent to those of the Morales Fm. or of the Santi Petri Fm. outcropping in their neighbourhoods. Moreover, the youngest Paleozoic Malaguide rocks have been dated here by means of conodonts (Navas-Parejo et al., 2012b). Before the present study, Upper Carboniferous (Bashkirian) deposits in the Malaguide Complex had been found only as shallow-water limestone pebbles and boulders within the Marbella Fm. (Buchroithner et al., 1980; Herbig, 1983, 1984). In the González site the coeval (Bashkirian) beds are represented by crinoid-rich hemipelagic limestones with conodonts. Therefore, this finding demonstrates the contemporary existence, somewhere in the Paleozoic Malaguide realms, of Bashkirian shallow marine and hemipelagic carbonate environments. According to their age, these Bashkirian sediments postdate deposition of the Carboniferous Almogía Fm. but predate deposition of the Marbella Fm., which was associated with the erosion of shallow water counterparts of the González Bashkirian conodont limestones.

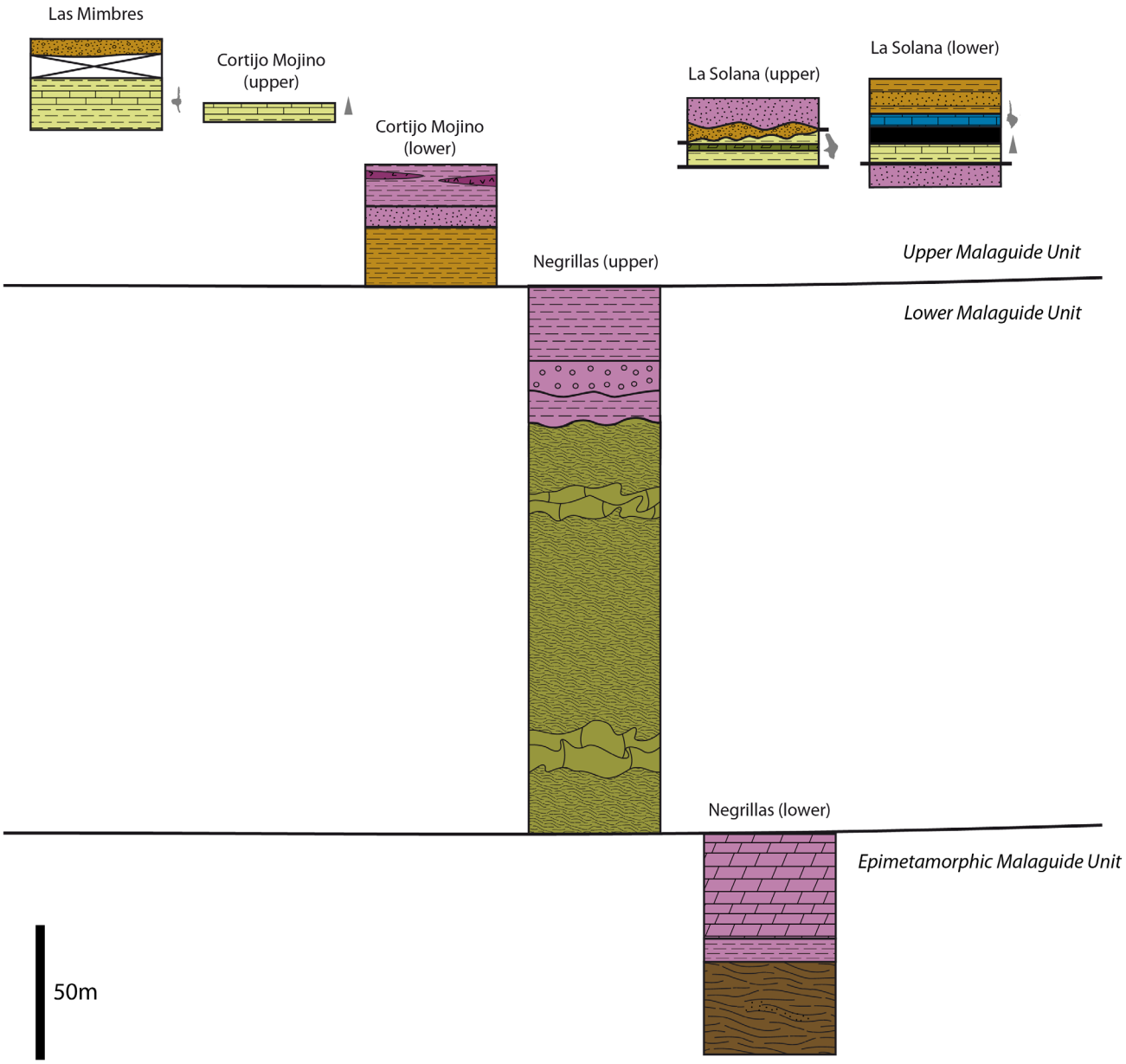
The unusual stratigraphic and tectonic features of the González outcrop suggest that it may be a tectonic remnant (klippe) of a wider Paleozoic succession that could be originally related to the successions outcropping today in the Upper Malaguide Unit. In the same way as the Ardales Castle Rock succession, which actually constitutes a Variscan upper thrust sheet ontop of a stratigraphically different Paleozoic succession at its footwall but both making part of the same Alpine Lower Malaguide Unit, the González outcrop could also be interpreted as a remnant of a pre-Alpine thust sheet within the Alpine Epimetamorphic Malaguide Unit.

II.4.4. Lower Malaguide Unit

The Alpine Lower Malaguide Unit is equivalent to the Cerro Colorado Unit of Roep and Mac Gillavry (1962) in the Vélez Rubio-Lorca Corridor, to the Unit 2 defined by García-Dueñas and Navarro-Vilá (1976) in the Cogollos Vega Zone, and to the Ardales-Mojón Unit of Martín-Algarra et al. (2009a, and b) in the Ardales area. In the latter area, however, only the Paleozoic succession of the Variscan Lower (Ermita) Subunit recognised within it is identical to that of the Alpine Lower Malaguide Units in the other areas. The following studied successsions belongs to this unit, from E to W: Fuente Alegre, Barranco Lobo, Taberno road, Pedrero and Cañada Hermosa/Rambla Bermeja within the Vélez Rubio-Lorca Corridor (Fig. II.4.1), Negrillas (upper) within the Cogollos Vega Zone (Fig. II.4.2), and Arroyo de la Torre and Arroyo Viñas (lower) in the Ardales region (Fig. II.4.3).

As stated above, the Lower Malaguide Unit is the largest (widest outcropping and thickest) Malaguide Unit in most areas, and equivalent terrains are the most widely present in the largest outcrops of the Malaguide Complex (Montes de Málaga and Costa del Sol). The standard stratigraphic succession of the Malaguide Complex actually corresponds to that of this unit, apart from the Marbella Conglomerate Fm., whose existence ontop of the Paleozoic succession of this unit cannot be totally excluded in the Montes de Málaga area.

The most typical stratigraphic feature of the Lower Malaguide Unit is the great thickness of its pre-Carboniferous succession, which is particularly characterized by the systematic presence of *Calizas alabeadas* facies within its stratigraphic succession. The *Calizas alabeadas* facies also constitute the most diagnostic (although not the exclusive) facies feature of the Santi Petri Fm., which never appears within the Alpine Upper Malaguide Unit. Nonetheless, some differences can be recognised between the succession of the Lower Malaguide Unit in the studied outcrops and the typical succession found in the Montes de Málaga area,



Malaguide Complex

Saladilla Formation (Permian?-Triassic)

- Dark dolomites (a) Reddish sandstones (b)
- Dark pelites (a) with basic volcanites (b)

Almogía Formation (Lower Carboniferous)

- Green-olive slates (a) and greywackes (b)
- Polymictic conglomerates (Retamares member)

Falcoña Formation (Lower Carboniferous)

- Visean bluish limestones
- Tournaisian lydites

Fossils

- Carboniferous conodonts
- Upper Devonian conodonts
- Lower Devonian conodonts
- Dacryoconarids

- Medium-bedded bluish limestones
- Silurian?-Devonian dark pelites and greywackes (a) with platy limestone intercalations (b)

Santi Petri Formation (Devonian?)

- Calcareous turbidites (*Calizas alabeadas*) (a) with pelitic intercalations (b)

Morales Formation (pre-Devonian?)

- Dark slates (a) and greywackes (b)

Figure II.4.2 . Stratigraphic correlation chart of the Malaguide Paleozoic in the Cogollos Vega Zone.

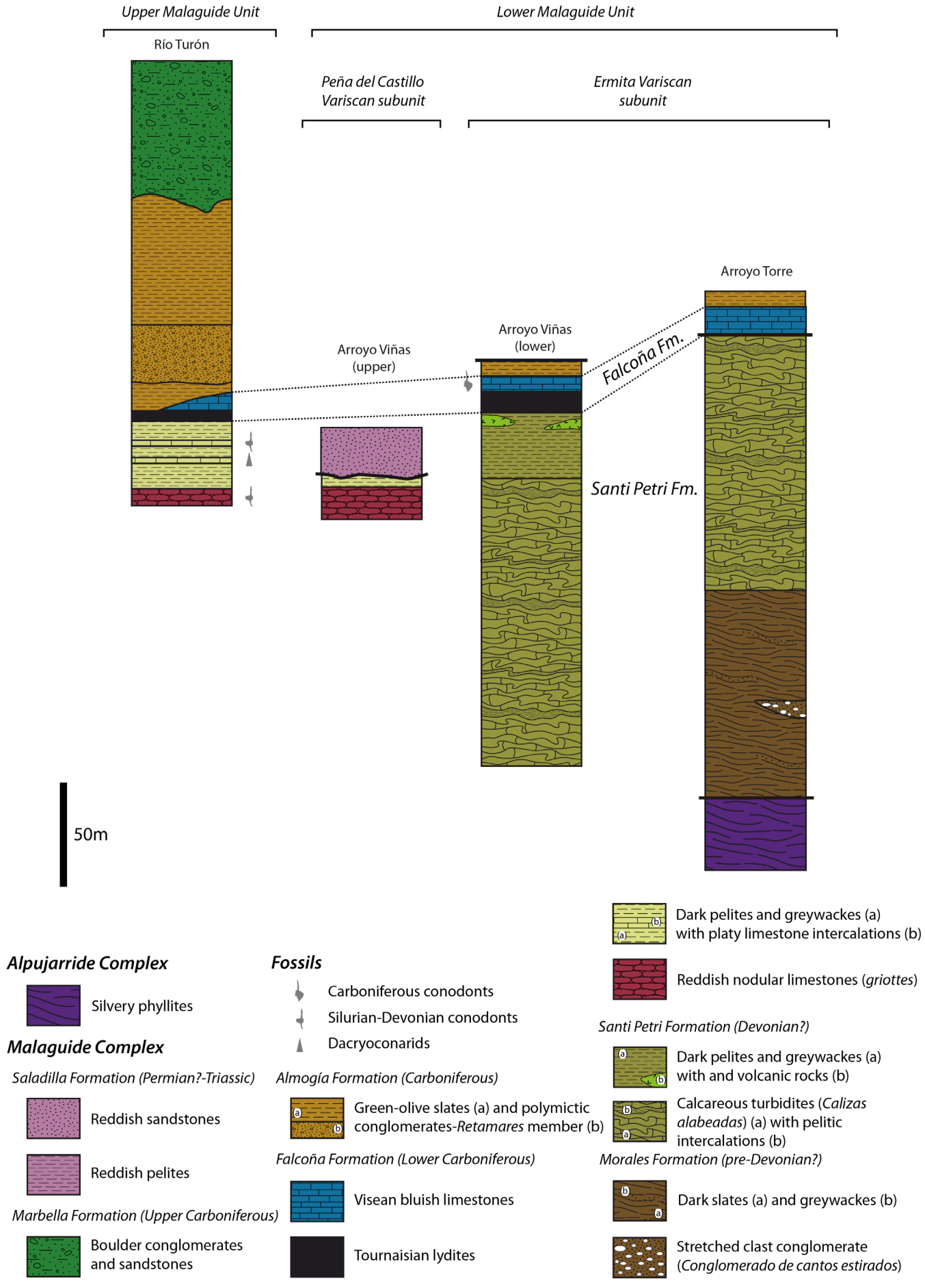


Figure II.4.3. Stratigraphic correlation chart of the Malaguide Paleozoic in the Ardales area.

which is usually much thicker.

In relation to the Morales Fm, it is thicker in the Vélez Rubio-Lorca Corridor (Fig. II.4.1 and thinner, or absent, in the other two areas (Figs. II.4.2 and II.4.3). This is certainly due to tectonic delamination of the tectonic base of this unit in the Malaguide outcrops close to the Internal-External zones boundary. For the same reason, the conglomeratic levels with stretched pebbles typical of the uppermost part of the Morales Fm. are rare in the studied outcrops of the Lower Malaguide Unit; they have been detected, however, in the Barranco Lobo to the east of Vélez Rubio and in the Rambla Bermeja to the south of Venta Quemada (westernmost part of the Velez Rubio Corridor: Fig. II.4.1). They have also been reported in the Arroyo Torre section to the south of Ardales (Fig. II.4.3). Another interesting feature of the Morales Fm. refers to the diabase dyke system that crosscut it. In the Lower Malaguide Unit it is exclusively present the Velez Rubio-Lorca Corridor (Fig. II.4.1) but not in the other two areas studied. The absence of dykes is a typical feature of the Lower Malaguide Unit in its central and western outcrops along the Internal-External domains boundary.

In the study areas important outcrops of the Calizas alabeadas facies of the Santi Petri Fm. are located in the Barranco Fuente Alegre area, and to the south of the Cañada de la Saladilla (to the east and southwest of Vélez Rubio respectively) and also in the Ardales region. In the Lower Malaguide Unit outcropping within the Cogollos Zone, the Santi Petri Fm. is much less thick than in the other areas due to tectonic causes, and the Calizas alabeadas facies are exclusively present there as thin intercalations within pelites and greywackes that are quite similar to those of the Morales Fm. (Fig. II.4.2). This kind of lateral variation within the Santi Petri Fm. (from an end-member richer in siliciclastic sediments, with a few isolated carbonate intercalations of calizas alabeadas facies, to another end-member where the classical Calizas alabeadas facies predominate, with only ocassional siliciclastic intercalations) is quite usual in other Malaguide outcrops of the Montes de Málaga, the Serranía de Ronda and the Costa del Sol. It probably reveals lateral variations in the efficiency of carbonate-rich distributary systems within the distal regions of the turbiditic basin in which the Santi Petri Fm was deposited.

Strictly speaking, the age of the Morales and Santi Petri Fms. remains unknown, as they have not provided any determinable fossil. A Devonian age is usually accepted for the Santi Petri Fm., because it stratigraphically underlies dated Lower Carboniferous beds of the Falcoña Fm. in different sectors. This Devonian hypothetical age is also supported by the presence of Emsian tentaculites-bearing limestones dated by conodonts on top of the pelite-sandstone interval above the Santi Petri Fm, at the foothills of the Ardales Castle Rock (top of Variscan Lower (Ermita) Subunit). It is unclear, however, if these tentaculites limestones constitute a true stratigraphic horizon of the Lower Malaguide Unit (both Alpine and Pre-Alpine) or a megaclast underlying the Falcoña Fm. lydites of this outcrop. Below the Santi Petri Fm., only isolated and undeterminable conodont debris were extracted from limestone clasts sampled from conglomerates in the upper part of the Morales Fm. (Barranco Lobo section). Further palaeontological studies on the limestones pebbles of the conglomerates with stretched clasts are needed to confirm the Devonian and pre-Devonian (Ordovician?-Silurian?) age of the Santi Petri and the Morales fms., respectively.

One yet unsolved mistery of the stratigraphy and palaeogeography of the Malaguide Complex, also related to the problem of the age of the Morales and Santi Petri Fms. is that of the provenance of the limestone block with Hirnantian conodonts obtained at the foothills of the Ardales Caste Rock, below the Falcoña Fm. lydites and immediately above the Emsian beds formerly mentioned, as well as that of its stratigraphic correlation to other Malaguide successions. Its provenance is unclear, as nothing equivalent has been found in any other site of the Lower Malaguide Unit. It neither can be excluded that it could have been eroded from stratigraphic successions equivalent to those of the Upper Malaguide Unit and redeposited within the sedimentary basin of the succession of the Lower Malaguide Unit in Devonian-Carboniferous time, before deposition of the Falcoña Fm. lydites. For analogous reasons, the stratigraphic successions of conodont-bearing Silurian-Devonian limestones of the Ardales Castle Rock and of the Arroyo de las Viñas (Fig. II.4.3) can not be easily correlated with the standard succession of the Lower Malaguide Unit, as they make part of a tectonically independent overlying unit (of Variscan age, indeed). On the contrary, they can be better correlated to the much more similar stratigraphic successions found in the Upper Malaguide Unit (Fig. II.4.3).

Overlying the Santi Petri Fm. the Falcoña Fm. occurs. It is nearly systematically constituted only by the black, thinly bedded and strongly deformed

radiolarites (lydites) of its lower member, whereas its upper limestone member is only observed in isolated scattered outcrops. The lydites have been recognized in the three studied areas, although clear stratigraphic relations with the underlying beds have not been recognised in the central sector (Cogollos Vega Zone): there, several typical lydite outcrops have been identified in different sites, but always strongly deformed and tectonically isolated from other under- or overlying Paleozoic formations. In the Arroyo de las Viñas section near Ardales the lydites overlie a pelite-sandstone interval that includes laterally discontinuous lenses of basic volcanic rocks that have not been found in the other areas (Fig. II.4.3). In the type section of the Falcoña Fm. Herbig (1983) mentioned the existence of similar volcanic rocks at the same stratigraphic location. Although the lydites usually are only a few metres thick, in the Barranco del Lobo section E of Vélez Rubio they reach several tens of metres and are much thicker than in other sections and sectors. These variations in thickness could be stratigraphic in origin, but more probably they are due to tectonic disturbances that have thinned the whole Falcoña Fm. in most areas. Within the Lower Malaguide Unit, the only horizons dated by means of conodonts in this study have been found in the upper member of the Falcoña Fm. (Pedrero and Fuente Alegre sections in the Vélez Rubio area and Arroyo Viñas to the southwest of Ardales: Fig. II.4.1).

The upper lithostratigraphic unit that appears in this Lower Malaguide Unit corresponds to the greenish pelites and greywackes of the Almogía Fm. Frequently they include conglomeratic beds of the Retamares member in many other isolated outcrops in the three studied areas.

II.4.5. Upper Malaguide Unit

The Upper Malaguide Unit includes both the Castillos and Salud units defined in the Vélez Rubio-Lorca Corridor by Roep and Mac Gillavry (1962), which have no differences regarding to the Paleozoic succession; the Unit 3 of García-Dueñas and Navarro-Vilá (1976) in the Cogollos Vega Zone, and the Castillejos-Turón Unit of Martín-Algarra et al. (2009a) in Ardales also make part of this unit. The following studied successsions belongs to the Upper Malaguide Unit, from E to W: El Piar, la Tejera and Los Gaticos within the Vélez Rubio Corridor (Fig. II.4.1), La Solana (lower and upper slice), Cortijo Mojino and Las Mimbres within the Cogollos Vega Zone (Fig. II.4.2), and Rio Turón in the Ardales region (Fig. II.4.3).

The Upper Malaguide Unit is mainly characterised by the presence of thin intervals (up to a few tens of metres thick at most) of Silurian-Devonian pelagic limestones bearing conodonts, dacryoconarids, crinoids and orthoceratids, among other macro and microfossils. Both members of the Falcoña Fm. make also part of this Upper Malaguide Unit, although Visean conodonts in the limestone member have been found only in the Cogollos Vega Zone (La Solana lower section, Fig. II.4.2). Fammenian conodonts have been found below the lydites of the Falcoña Fm. in the Vélez Rubio-Lorca Corridor (Gaticos section Fig. II.4.1). The green-olive slates, greywackes and polymictic conglomerates of the Almogía Fm. and, especially, the Marbella Fm. are very well developed in this unit.

Neither the Morales Fm. nor the Santi Petri Fm have been recognised in this unit, since slates and pelites found there are systematically related either to Silurian-Devonian pelagic (not turbiditic) limestone intercalations or to the Carboniferous Retamares member. As stated above, the Silurian-Devonian limestones with conodonts of the Arroyo Viñas (Fig. II.4.3) and their laterally equivalent reddish nodular limestones with orthoceras of the Ardales Castle Rock succession (Kockel, 1959; Rodríguez-Cañero, 1993), which make part of the Upper Variscan Subunit of the Lower Malaguide Unit (Martín-Algarra et al., 2009a and b), have the same facies and age than the Silurian-Lower Devonian nodular limestone beds and dacryoconarid limestones of the Río Turon section. These are also equivalent to those found and dated by means of conodonts in the Vélez Rubio area by van den Boogaard (1965), which also make part of the Upper Malaguide Unit.

Within the pre-Carboniferous stratigraphic sequence of the Upper Malaguide Unit, in the stratigraphically equivalent Peña del Castillo Variscan Subunit and in the González outcrop different facies have been recognised, among them: i) reddish nodular limestones with orthocerathids

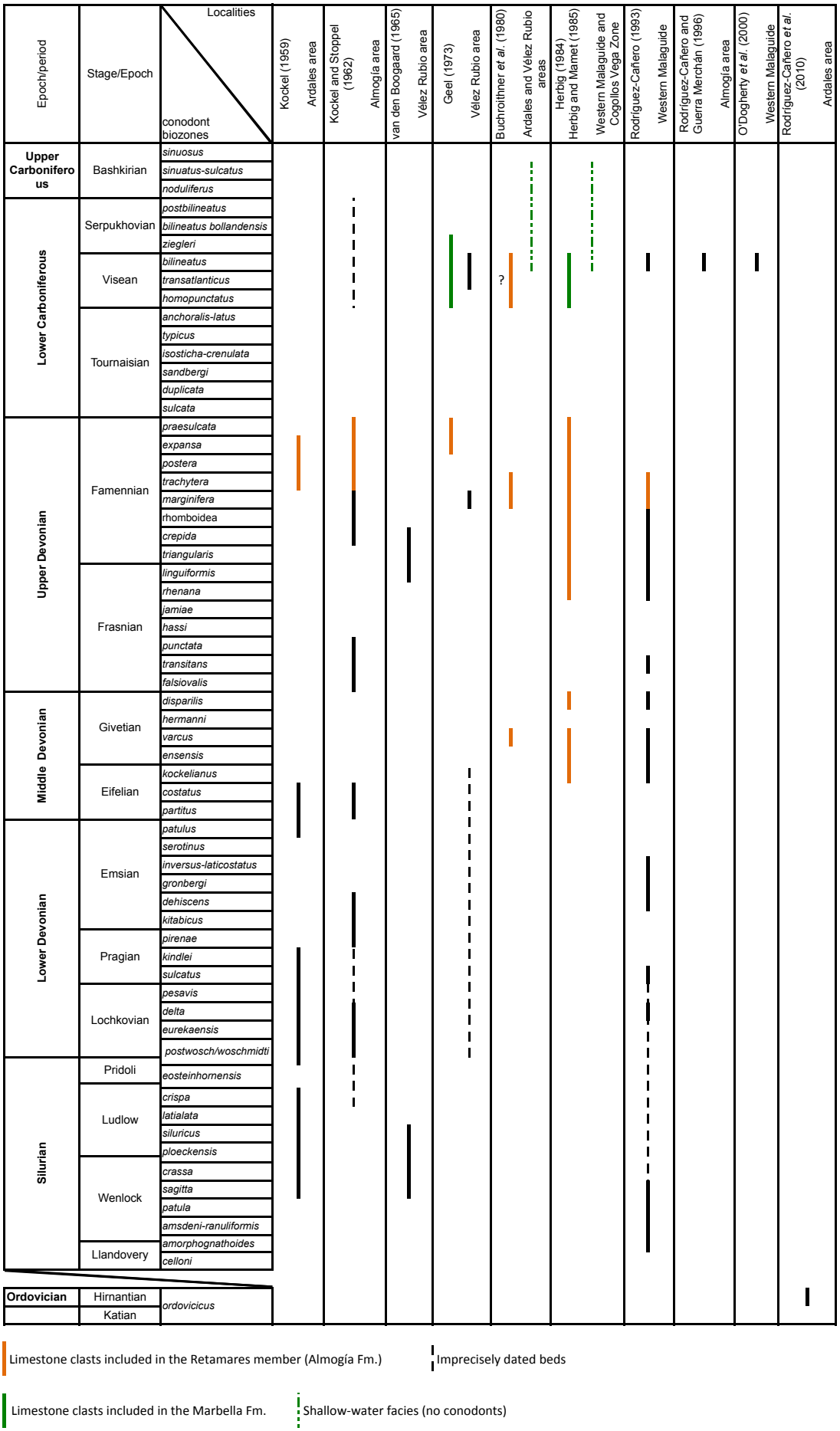


Figure II.4.4. Previous biostratigraphic data for the Malaguide Paleozoic (by means of conodonts except for the Marbella Fm.).

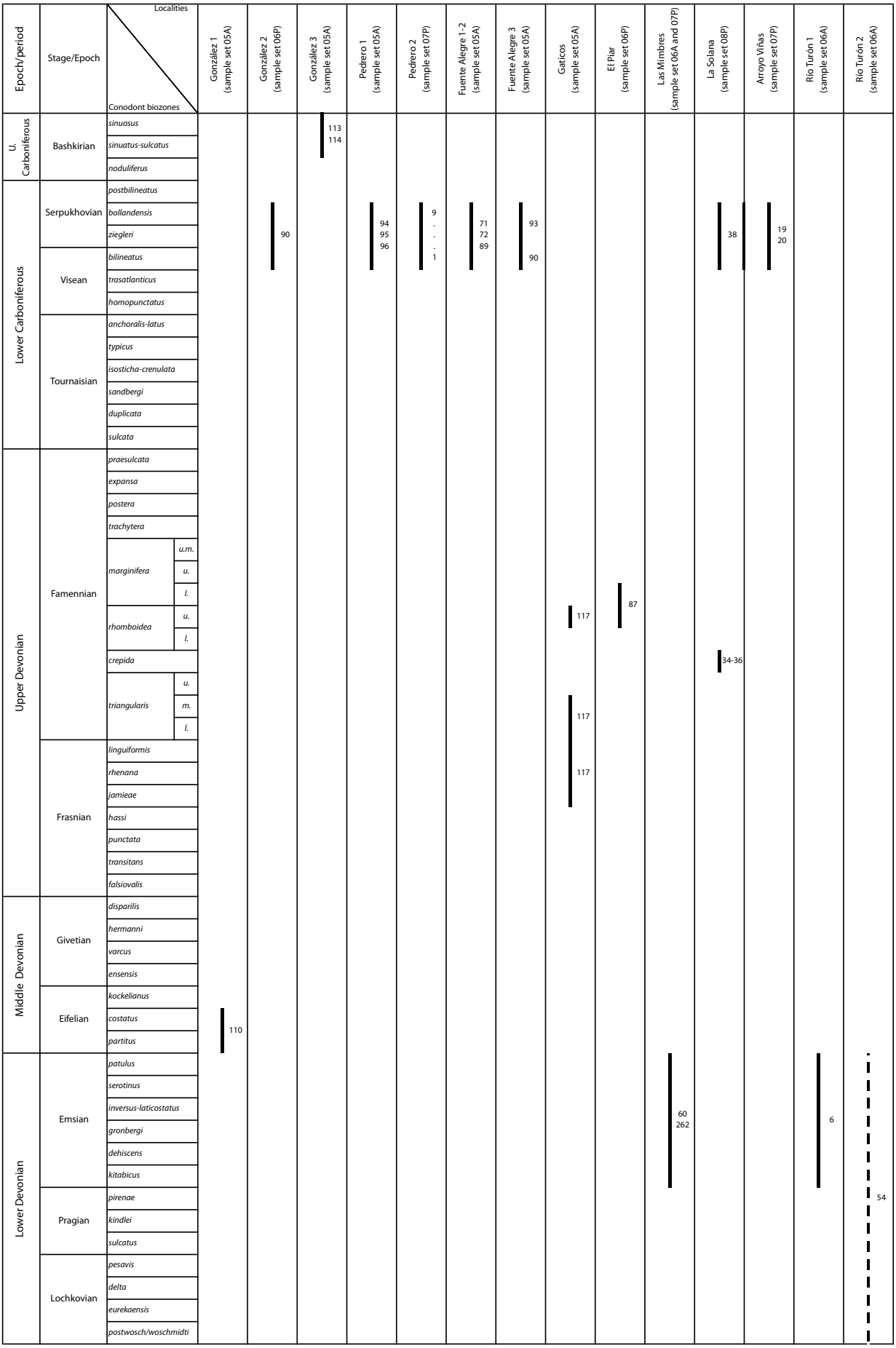


Figure II.4.5. Biostratigraphic results in the Paleozoic of the Malaguide Complex.

(Silurian-Lower Devonian, Arroyo Viñas and Río Turón sections: Fig. II.4.3); ii) greyish-bluish massive limestones (Lower Devonian, Las Mimbres section: Fig. II. 4.2), iii) strongly recrystallized and marbly limestones (Middle Devonian, González 1 section); iv) well bedded greyish-bluish to greyish-reddish and sometimes nodular limestones with dacryoconariids (classical Lower-Middle Devonian tentaculites limestones, Mojino section: Fig. II.4.2; Arroyo de las Viñas: Fig. II.4.3); v) well bedded to platy yellowish limestones with dacryoconarids alternating with calcareous and/or varicoloured pelites (Lower-Middle Devonian platy tentaculites limestones, La Solana Lower section: Fig. II.4.2); and vi) bluish micritic limestones (Upper Devonian, Gaticos and la Solana upper sections: figs. II.4.1 and II.4.2). Equivalent Silurian and especially Devonian successions have been also recognized in many other outcrops in the Montes de Málaga (Almogía area), Ardales and Costa del Sol by previous authors (Kockel, 1959; 1960; Kockel and Stoppel, 1962; Herbig, 1985; Rodríguez-Cañero, 1993, 1996; see Fig. II.4.4). In some places, an Upper Devonian breccia level with pebbly mudstone facies, probably corresponding to a Famennian olistostrome) has been identified underlying the Falcoña Fm. (Falcoña outcrop: O'Dogherty et al., 2001; Gaticos section: Fig. II.4.1).

As stated above, in the Upper Malaguide Unit the Falcoña and Almogía fms. do not differ essentially from the same formations in the underlying units. The Lower Carboniferous age of the Falcoña Fm is here constrained by conodonts in the upper limestone member only in La Solana section and by the presence of Fammenian beds below the lydites of Los Gaticos section. The Carboniferous age of the Almogia Fm. in the Upper Malaguide Unit of the Vélez Rubio area is well constrained by the presence of limestone clasts with Devonian conodonts (Piar section), in addition to the common presence of lydite clasts. But one of the most impressive feature of the Upper Malaguide Unit is the widespread occurrence of the Marbella Fm. Large outcrops of it have been observed in the three studied sectors, although this formation has not been studied in detail.

II.5. An outline of the Malaguide Paleozoic palaeogeographic evolution

In relation to the sedimentary and palaeogeographic significance of the Malaguide stratigraphy, the pre-Carboniferous record points out to lateral facies changes in a sedimentary basin with two contemporaneous (distal and proximal) depositional areas and slightly different tectono-stratigraphic evolutions (Fig. II.5.1). One of these areas is typified by the classical Malaguide Paleozoic succession that characterizes the Lower Malaguide Unit (appart from the Upper Variscan Unit represented by the Ardales Castle Rock and equivalent successions) and most of the Epimetamorphic Malaguide Unit (appart from the Cortijo de los González succession). It represents a distal, basinal depositional context. Another area is that where the Silurian-Devonian stratigraphic succession of pelagic limestones and pelites of the Upper Malaguide Unit, the successions of the Upper Variscan Subunit Ardales Castle Rock and equivalents, and probably also the González succession, were deposited. It represents an also yet relatively distal open marine area within a continental margin, but much less deep and much less distal than the former (then proximal, in comparison to it). Silurian-Devonian deposition in this proximal area of the basin is typical of moderately deep and not turbiditic hemipelagic environments, and/or of pelagic high bottom areas with pelagic condensed carbonate sedimentation. In addition, indirect evidence of the coeval existence of shallow marine proximal environments also exists, as some clasts showing carbonate platform facies have been found just below the Falcoña Fm. lydites (Falcoña, O'Dogherty et al., 2001; Fuente Alegre section, this study). Nonetheless, the existence of platform areas was much better documented by Herbig (1984, and related papers) for Carboniferous times from Visean to Bashkirian carbonate clasts in the Late Carboniferous conglomerates of the Marbella Fm.

The classical stratigraphy of the Lower Malaguide Unit records a basinal sedimentary succession because it is dominated by fine-grained terrigenous-clastics (with very rare, thin and laterally discontinuous coarser grained intervals). It was deposited in moderately deep to very deep marine, predominantly turbiditic sedimentary environments from pre-Silurian to Early Carboniferous times. This points to a starting original depositional site for this succession in marine environments, as no evidence of continental deposition exists anywhere at the stratigraphic base of the Malaguide successions. Upwards, the sedimentary features indicate base of the slope to basin plain depositional settings, then in the distal region of a continental margin.

The stratigraphic succession from Morales to Santi Petri to Falcoña formations clearly indicates a deepening upward evolution: sedimentation started in pre-Devonian times (Cambro-Ordovician?) with siliciclastic deposition, in probably moderately deep marine environments, as clear evidences of turbiditic sedimentation are rare in the Morales Fm., although this could be due to the strong tectonic deformation that affects this succession). The sedimentation evolved to distal turbiditic environments during deposition of the Santi Petri Fm. Finally deposition accounted in very deep marine, starved, basinal environments (Falcoña Fm.).

The siliciclastic nature of the starting Malaguide Paleozoic sedimentation (Morales Fm.) was probably due to the high latitudinal location of the European regions of Northern Gondwana, from which the Malaguide realms also made part, in Cambrian to earliest Silurian time. Carbonate sedimentation in such cold or cool-water settings near to the South Pole should be certainly hampered, and siliciclastic deposition favoured. The upward evolution of the stratigraphic succession indicates a gradual change to warmer depositional environments with ocassional carbonate sedimentation. This represents the local record of the northwards drift of the Malaguide realms, as a part of the Northern Gondwanic regions, towards lower paleolatitudes.

The fine-grained character, in general, of the pelites and psammites of the Morales Fm.

suggests deposition in sedimentary environments of moderate to low hydrodynamic energy, probably in not extremely deep but relatively distal marine environments of a passive margin. Nevertheless, from time to time the basin also received some coarse-grained siliciclatic contribution, sometimes bearing rare carbonate clasts (Barranco Lobo), especially near the top of the Morales Fm. (horizon of conglomerates with stretched clasts).

During (Silurian?-)Devonian, the sedimentation rate probably increased, as the Santi Petri Fm. reaches frequently several hundred of metres in thickness. But this sedimentation certainly did not compensate subsidence, as the depositional environment gradually becomes deeper and deeper. The deposition of the pelites, greywackes and calcareous turbidites of the Santi Petri Fm. indicates mixed siliciclastic-carbonatic distal turbiditic sedimentation corresponding to distal slope and basinal turbiditic environments. Consequently, at the end of deposition of the Santi Petri Fm., the sedimentary context was that typical of a mature divergent continental margin.

During the same time, in proximal zones of the basin, comparatively thin and moderately deep pelagic carbonate sediments were being deposited. The Silurian (frequently reddish) and lowermost Devonian (commonly bluish to, less frequently, bluish-reddish or pinkish) nodular limestones represent condensed deposits on top of high bottom pelagic areas. These are typical of the early geodynamic evolution of marine domains of continental margins. Their deposition usually accounts on comparatively rised marine bottoms on rifted and sometimes tilted blocks generated during the early evolutionary stages of divergent continental margins. During the Devonian the sedimentation in these areas was dominated by pelagic bluish to grayish pelagic limestones, usually bearing dacryoconarids. Either if more or less massive of if thinly to very thinly bedded and alternating with more or less calcareous pelites and very fine-grained sandstones, these facies points to a gradual deepening of the sedimentary environments.

The maximum depth in the basin was reached during the earliest Carboniferous (Tournaisian). It is testified by the widespread deposition of the lower radiolaritic member of the Falcoña Fm. The transition to the Falcoña Fm. was, in both types of successions, gradual, especially in the most basinal succession is represented by varicoloured pelites, frequently rich in Mn, that alternate with fine-grained siliciclastic and fine to coarse-grained carbonatic turbidites, and with radiolarites. Some tectonic instability of the basin during this transition is recorded by the local intercalation of basic volcanics (Arroyo Viñas section) or by the carbonate turbidites themselves, especially where they are coarse-grained (Fuente Alegre section) or are related to deposition of carbonate breccias (Gaticos section). The latter case suggests a sudden drowning of the Fammenian carbonate environments to much deeper settings where accounted the deposition of the siliceous (radiolaritic) deposits of the Falcoña Fm. Later, during the Visean epoch, the basin became gradually shallower or, more probably, a sudden and widespread deepening event of the Carbonate Compensation Depth (CCD) accounted, as the radiolaritic sedimentation was rapidly replaced by deep marine but carbonatic pelagic sedimentation at that time in many Paleozoic regions around Europe and North Africa.

A sharp change in sedimentary conditions accounted in Serpukhovian time, as demonstrated by the widespread deposition of the thick and widespread coarse-grained Culm-like facies of the Almogía Fm. This rapid change in the sedimentary regime suggests an important and geologically rapid change in the geodynamic setting of the basin, that evolved from a mature divergent to a starting convergent continental margin. Rising reliefs in proximal areas of the margin were rapidly eroded and supplied great amounts of coarse-grained immature siliclastic detritus to the adjoining yet deep marine basin. In some areas, the transition from the underlying Falcoña Fm. to the overlying (fine-grained, in this case) siliciclastic deposits was gradual. However, in other areas, it is marked by an unconformity, although the stratigraphic relations in many of them are strongly disturbed tectonically, due to the pre-Alpine and/or the Alpine orogenic evolution. Such unconformable relations explain why the carbonate member of the Falcoña Fm., or even the whole formation, is absent in many areas where the Santi Petri and Almogía fms. are in direct contact.

The deposition of the channeled Retamares conglomerates and coarse-grained sandstones indicates deposition in proximal deep sea fan turbiditic environments. The presence of clasts of lydites, and of Devonian limestones in the Retamares member (figs. II.4.4 and II.4.5) point out to canibalistic erosional-depositional processes related to the early orogenic Variscan evolution in wide parts of

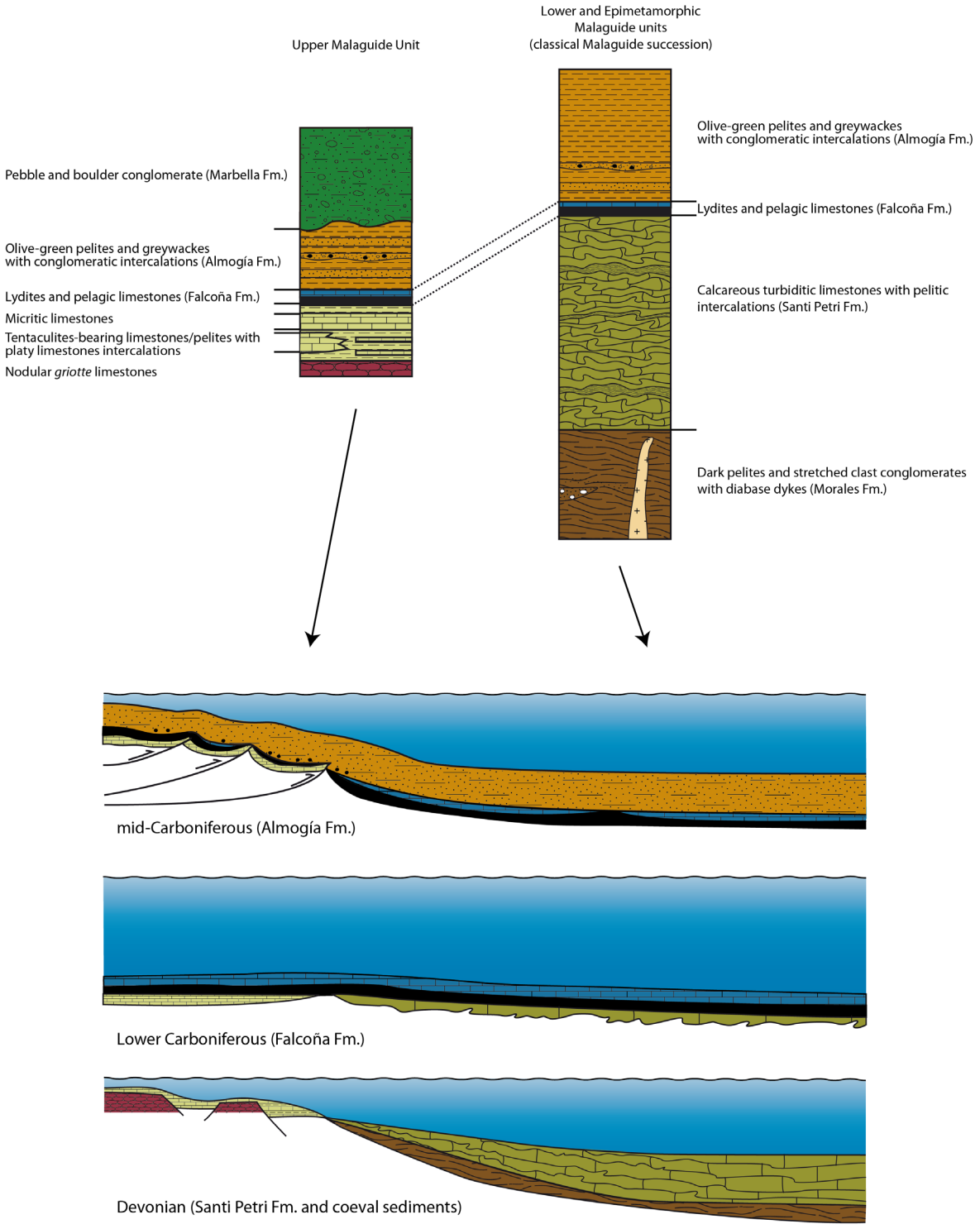


Figure II.5.1. Synthetic stratigraphic columns for the two distinguished Malaguide Paleozoic successions and proposed palaeogeographical model.

the Malaguide basin. The tectonic deformation and rising of reliefs in proximal areas was certainly very important, as large amounts of terrigenous detritus, and large clasts of magmatic and metamorphic rocks (granites, gneisses, schists...) supplied to the Almogía basin testify for a deep erosion not only of sediments previously deposited in the sedimentary basin but also of the continental basement rocks of the proximal region of the margin. Among the eroded clasts there also appear carbonate cobbles and boulders eroded from Visean shallow marine carbonate platforms whose deposition was coeval to pelagic sedimentation (upper member of the Falcoña Fm.) in more distal basinal settings of the margin.

The gradual decrease in the tectonic unstability of the basin during the deposition of the Almogía Fm. is clearly evidenced by its fining-and thinning-upwards sequential evolution. This interpretation is reinforced by the finding of Bashkirian crinoid-bearing conodonts limestones in the González 3 section. Its presence clearly documents that a lateral evolution from a shallow marine carbonate platforms to carbonate hemipelagic to pelagic environments, comparable to that above mentioned for the Viséan-Serpukhovian interval, also accounted in Bashkirian-post-Bashkirian times. It also demostrates that the Carboniferous siliciclastic sedimentation was punctuated, at least in some areas of the basin, by carbonate deposition at the beginning of the Late Carboniferous, and that this precise time was a period of relative tectonic quiescence in the basin. After that, a renewed and probably stronger orogenic pulse destroyed the Bashkirian carbonate platform eroding enormous clast of them together with large clasts of granites and gneises from the continental crusts substratum, and deposited them by debris flows in the adjoining deep marine basin, to form the Marbella Fm. Immediately after, the basin was totally closed and transformed into a segment of the just born Variscan Orogen.

CHAPTER III

Paleozoic of the Southern Calabria-Peloritani Terrane

III. 1. The Stilo Unit in the Serre Massif
III. 2. The Stilo Unit in the Aspromonte Massif
III. 3. Synthesis of the Stilo Unit Paleozoic stratigraphy
III. 4. The Longi-Taormina Unit
III. 5. Paleozoic tectono-stratigraphic evolution of the Longi-Taormina Unit

III.1.1. Previous stratigraphic studies in the Serre Massif

The Stilo Unit is the highest thrust-sheet of the Calabrian nappe stack and crops out discontinuously only in southern Calabria, from the Serre Massif to the Aspromonte (Fig. III.1.1.A). This unit is made up of a Meso-Cenozoic cover and a pre-Mesozoic basement composed of Carboniferous granites and pre-Devonian (Cambrian-Ordovician?)-Carboniferous metamorphites.

The most complete outcrop of the metasedimentary Stilo Paleozoic succession is located in the Serre Massif, in the surroundings of Bivongi and Stilo (Fig. III.1.1.B). It is known as the Bivongi sequence (Majesté-Menjoulas *et al.*, 1984; Bouillin *et al.*, 1987). This succession was firstly described by Knoche (1977, 1978) in the Torrente Crocco, where he established a lithologic succession made up of alternations of greywackes,

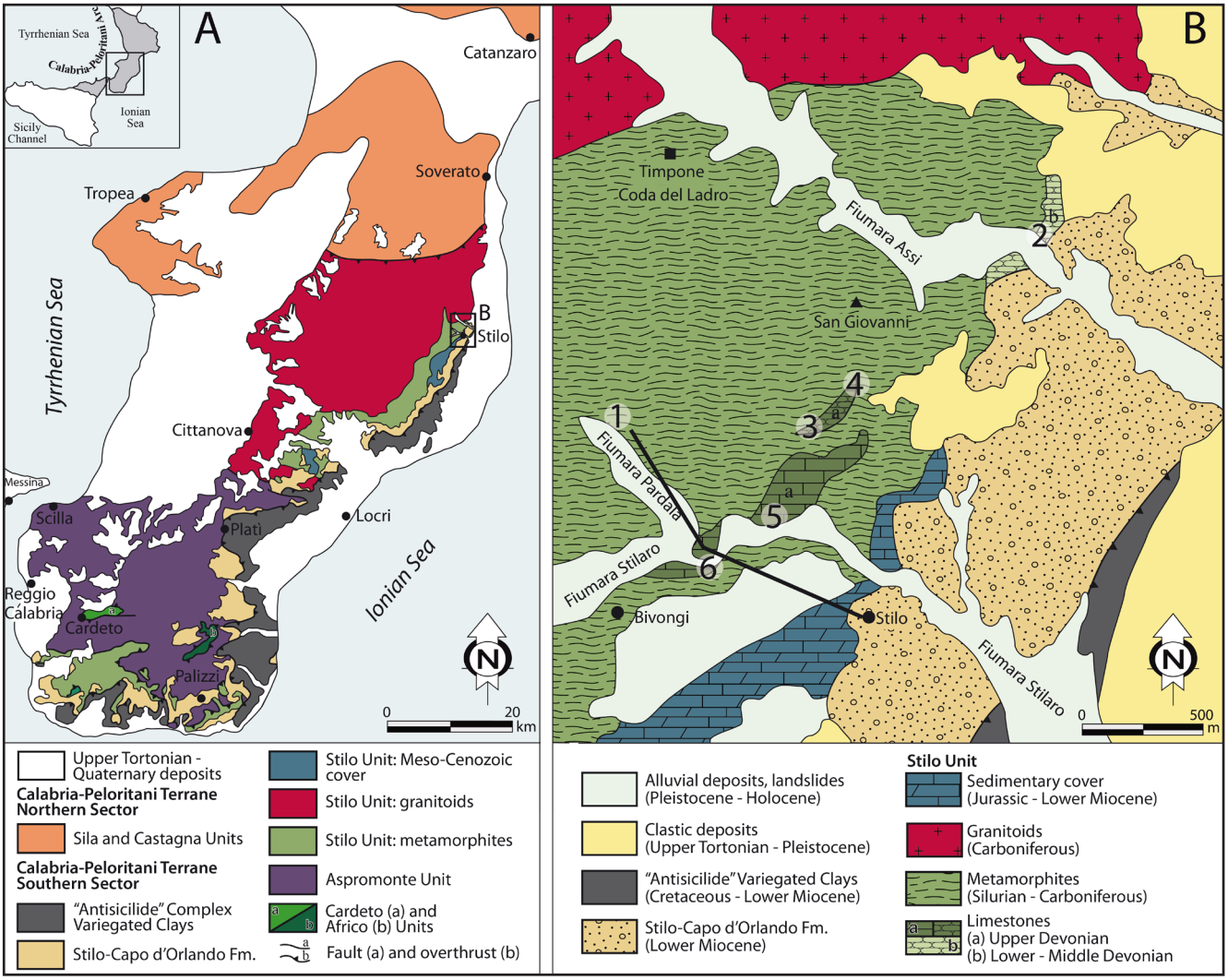


Figure III.1.1. ***A.*** *Tectonic sketch of the Southern Sector of the Calabria-Peloritani Terrane in Calabria.* ***B.*** *Geological sketch of the Stilo-Bivongi area with location of the studied sections:* ***1.*** *Pardala;* ***2.*** *Assi;* ***3.*** *S. Giovanni 1;* ***4.*** *S. Giovanni 2;* ***5.*** *Stilaro;* ***6.*** *Ponte Vina.*

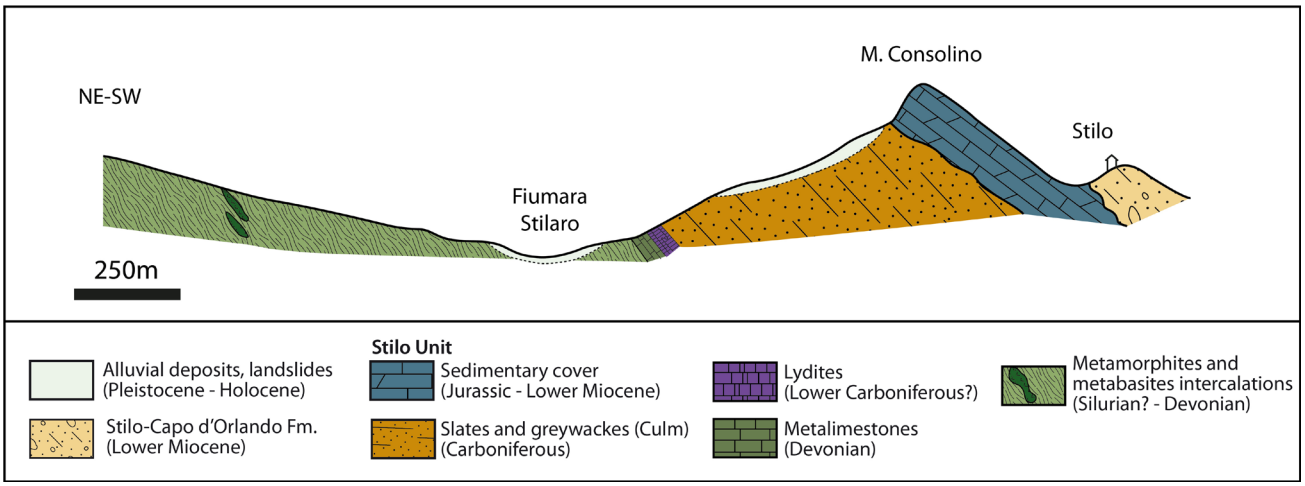


Figure III.1.2. General geological cross section in the Bivongi area (see location in Fig. III.1.1).

cherts, schists, limestones and, in the topmost part, crinoid-bearing limestones. He stated that this sedimentary sequence was deposited in a basin, probably turbiditic, with frequent detritic influx.

Majesté-Menjoulas *et al*. (1984) studied the Paleozoic of the Stilo Unit in the same area, mostly along the road to San Giovanni Vecchio and the ravines that crosscut this Paleozoic succession between the Fiumara Stilaro and the Fiumara Assi. They distinguished the following lithostratigraphic units, from base to top:

1. Lower detritic formation: grey-greenish sandstones and slates with intercalated basic rocks and pillow lavas (Pardalá stream). They stated that this formation was affected by contact metamorphism of the Hercynian granite of Stilo.

2. Black graphitic slates (ampelites) and black silicic slates.

3.a. Very thin-bedded grey limestones.

3.b. Very thick detritic formation, mainly slates and thin sandstone intercalations. This formation is crosscut by numerous stretched and boudinated veins of quartz. An intercalated and boudinated limestone bed yielded crinoids and bryozoan fragments.

3.c. Nodular limestones constituted by nodules of grey, red, violet or green micritic limestones. They show the typical Devonian facies of the SW of France (griotte limestones).

4.a. Black silicic rocks with radiolarians, rich in organic matter (lydites) a few of metres thick. In the Crocco stream, thin greenish vulcano-sedimentary beds are intercalated.

4.b. Upper detritic sequence. Black pelites and grey (yellowish when weathered) turbidites intercalations. Locally they include plant remains.

In the absence of other dating methods, Majesté-Menjoulas *et al*. (1984) proposed different ages to these diverse lithologic groups, based on their resemblance with the stratigraphic sequences mainly of the Pyrenees, but also of the Montagne Noire, Sardinia, Kabylia and the Rif:

1. They pointed out that their facies were analogous to those dated as Cambro-Ordovician in the Sila Massif by means of acritarchs (Bouillin *et al*., 1984); therefore, they should be Ordovician.

2. This term show the typical Silurian features of the Pyrenees, the Montagne Noire and several sequences in Sardinia.

3.a. This interval reminds, due to its facies and stratigraphic position, to the *calcaires à laminites* of the Lower Devonian of the Pyrenees.

3.b. In the crinoid-bearing limestones, they found fragments of conodonts, which remain Lower-Middle Devonian forms.

3.c. They dated this unit as Famennian by means of conodonts and correlated it with coeval beds with similar facies in the Pyrenees and the Montagne Noire. In these regions, such nodular limestones underlie lydites, in the same way as in the Bivongi sequence. The interval 4.a. should then be lowermost Carboniferous.

4.b. These facies remind the Culm-type

Carboniferous series known in the Pyrenees, the Kabylies and the Rif.

Bouillin *et al*. (1987) studied four Paleozoic successions belonging to the Calabria-Peloritani Terrane: in the Sila Massif (Laurenzana formation), in the Bivongi-Pazzano area (Bivongi sequence), in the Aspromonte (Pietrapennata sequence) and in the Peloritains Mts. They slightly modified the stratigraphic frame previously proposed by themselves (Majesté-Mejoulas *et al*., 1984) for the Bivongi sequence, mainly in the following:

2.b. In this unit, previously interpreted as Lower Devonian by facies resemblance with the Paleozoic of the Central Pyrenees, they found conodont fragments with Late Silurian affinity.

3.a. There, they found conodont fragments of probable Middle Devonian age.

3.c. They confirmed a Famennian age for these griotte facies by means of conodonts.

III.1.2. Previous biostratigraphic data

Cambro-Ordovician, Silurian? and Devonian fossils have been found in the Paleozoic succession of the Stilo Unit, especially in the Bivongi-Guardavalle area. The first fossil discovered in the Paleozoic of the Stilo-Pazzano area was published by Franco (1881). He found a trilobite cephalon of *Phacops laevis* Münster, which allowed to infer a Devonian age.

As part of a preliminary study, Görler and Ibbeken (1970) described Paleozoic fossils in the Fiumara Assi, the Fiumara Stilaro and the Torrente Crocco. They found corals, bryozoans, crinoids, tentaculites, orthoceras, planispirals cephalopods, trilobites and ostracods, all of them very badly preserved. In this preliminary study, they assigned a Middle Devonian age to such rocks, although they did not discard a Silurian, a Lower or an Upper Devonian age. They announced further conodont studies that have never been published.

Afchain (1970) found dacryoconarids in the limestone intercalations within the metamorphites, which confirmed the Devonian age previously dated by Franco (1881). In the same year, de Capoa-Bonardi (1970) found *Palmatolepis perlobata perlobata* Ulrich and Bassler, a lower Famennian (Upper Devonian) conodont species, near the Ponte Vina. This section has been resampled and studied for this dissertation (see section III.1.4 for biostratigraphic details) and the preliminary results have been already published (Navas-Parejo *et al*., 2009b; see appendix I).

Majesté-Mejoulas *et al*. (1984) studied the Bivongi-Guardavalle area, looking for acritarchs, spores and conodonts but they only found identifiable conodonts in the *griotte* limestones (see previous section). They found Famennian conodont associations belonging to three biostratigraphic intervals: upper *crepida*-upper *rhomboidea* zones, *marginifera* zone, and upper *velifer*-lower *costatus* zones. Moreover, in the crinoid-bearing limestones (their interval 3.b, see previous section) they found fragments of conodonts with a likely Lower-Middle Devonian age.

Bouillin *et al.* (1987) also studied the Bivongi sequence and they found conodonts in several limestone beds. In the thin-bedded limestones of their interval 2.b. (see previous section), they found conodont fragments that remind Upper Silurian species. In the slates with very thin-bedded sandstone intercalations (their interval 3.b.), a limestone bed yielded fragments of crinoids, bryozoans and conodonts which allowed them to consider a Middle Devonian age. They also confirmed the Famennian age for the *griotte* limestones by the presence of several *Polygnathus* and *Palmatolepis* species.

III.1.3. Studied sections

Eight sections have been studied, measured and sampled for conodonts in order to establish the stratigraphic succession of the Paleozoic of the Stilo Unit in Calabria (see location in Fig. III.1.1.B). Two of them were made in the Aspromonte Massif and they will be described in the next section, whereas the litho and biostratigraphic results obtained in the six sections studied in the Serre Massif (Stilo-

Bivongi area) are presented in the following lines.

In the Bivongi-Stilo area, except for the Pardalà section, five sections (Assi, San Giovanni 1 and 2, Ponte Vina and Stilaro), have released identifiable conodonts, and several levels of Devonian age have been characterized. The biostratigraphic results of two of these sections (Assi and Ponte Vina) have been already published during the course of this dissertation (Navas-Parejo *et al*. 2009a and b, see appendix I). These results, as well as new data from these sections are also summarized below.

Pardalà section

The lowest beds of the Stilo Unit are those outcropping along a unpaved track that starts from the road SP 95 about 500 m to the NW of the Ponte Vina, and that runs parallel to the Fiumara Pardalà. There, a several hundreds of metres thick metapelitic succession crops out, dipping to the south (1 in Fig. III.1.1.B; Fig. III.1.3, Fig. III.1.4a). This is here considered as the lower part of the Bivongi Paleozoic sequence, and begins with a few hundreds metres of alternating, gray to greenish greywackes and slates, with intercalations of metabasites (Fig. III.1.4b) that gradually disappear upwards in the succession. Around one hundred of metres above the last metabasite intercalation, the Pardalà metapelites, with silvery-dark gray colours, intercalate a laterally discontinuous bed, up to 1m thick, of strongly foliated grey limestones (Fig. III.1.4.c). Five samples have been taken in this interval but, unfortunately, none of them have yielded conodonts. A few tens of metres upwards in the succession, within a partially covered area, the metapelites intercalate a few metres thick black interval, constituted by silicic black shales and black clays rich in organic matter (Fig. III.1.4.d). Therefore, the presence of ampelites in the Fiumara Pardalà, first mentioned by Majesté-Menjoulas *et al.* (1984) and by Bouillin *et al.* (1987), has been confirmed. Towards the top of the succession and intercalated within the metapelites some laterally discontinuous (probably due to tectonic disturbances and soil cover), slightly nodular and thin bedded limestone beds appear (Figs. III.1.4.e,f). These limestone lenses are those present in the other five studied sections.

Assi section

To the NE of San Giovanni, the Stilo Unit widely crops out on both sides of the Fiumara Assi, but it is particularly well exposed on the left versant, where the Assi succession was studied (2 in Fig. III.1.1.B; Fig. III.1.5). Structural and biostratigraphic data demonstrate that the Assi succession is overturned (Navas-Parejo *et al*., 2009a, see appendix I and section III.1.4 for biostratigraphic details). The stratigraphically lowermost beds of the Assi succession lie directly under the basal conglomerates of the Stilo-Capo d'Orlando Fm. (Fig. III.1.6.a). They are formed by a few metres of thin- to medium-bedded, parallel- laminated and partly nodular limestones, alternating with yellowish and greyish-black fine-grained limestones, with some orthoceratids. This interval has released several Lochkovian-Lower Pragian conodonts (Lower Devonian, see section III.1.4 for details). These carbonate layers are followed by a more pelitic

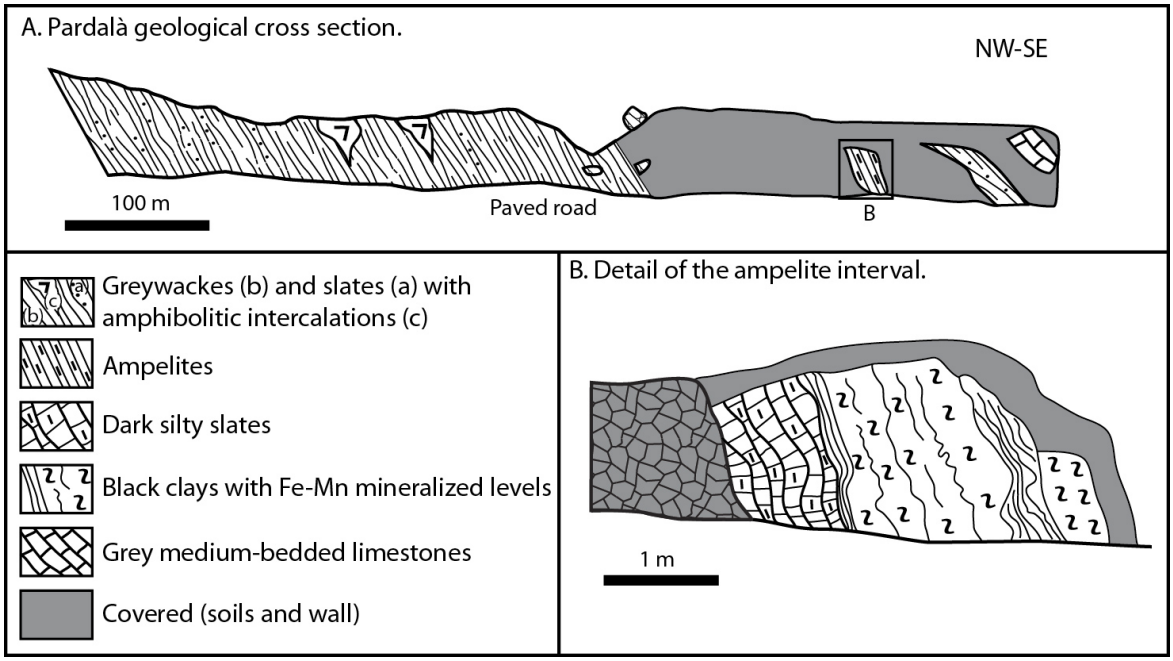


Figure III.1.3. Geological cross section in the Torrente Pardalà.

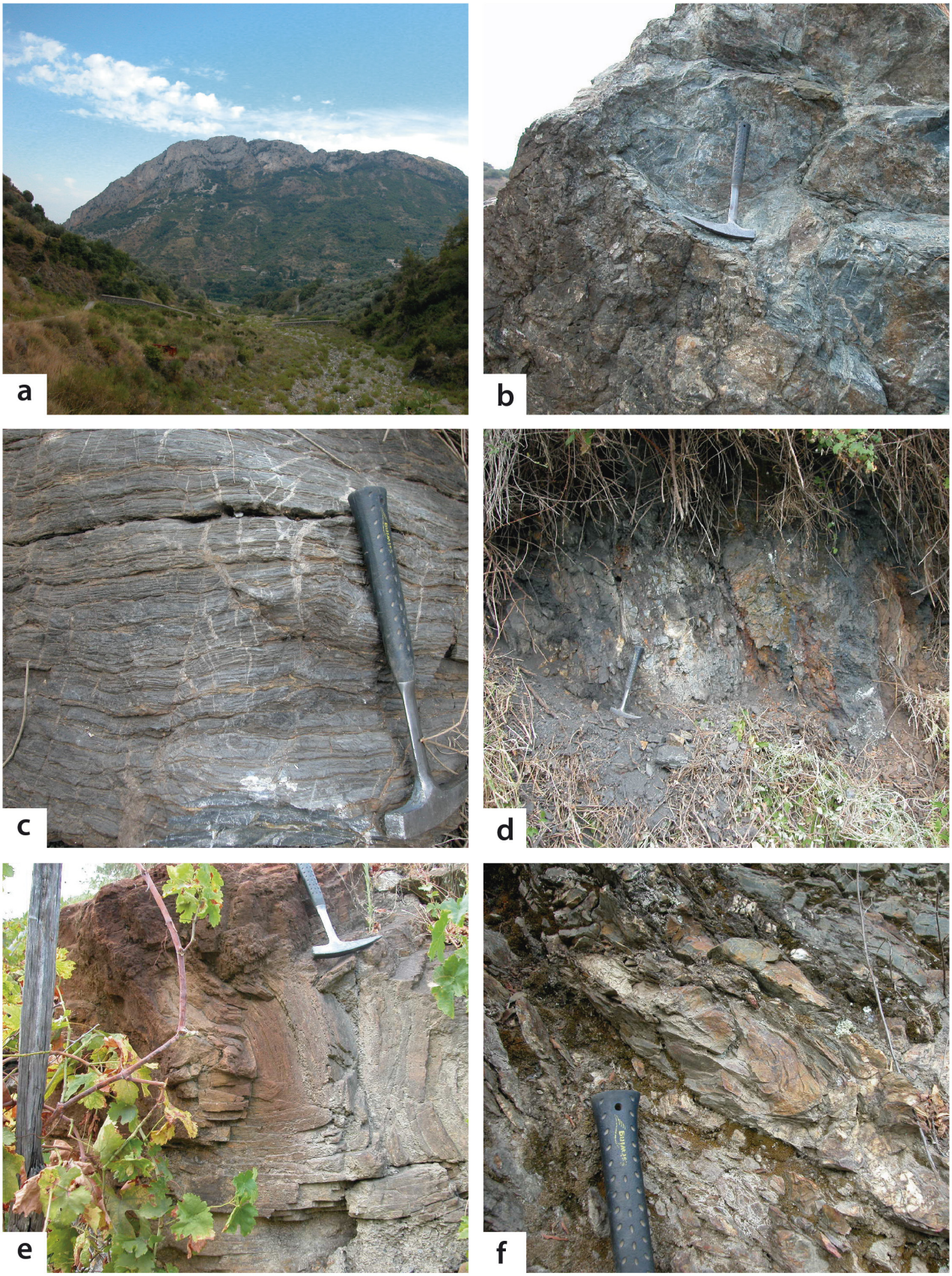


Figure III.1.4. Pardalà section. ***a,*** *M. Consolino view from the Torrente Pardalà (from the NE) in the background and part of the Pardalà section to the left;* ***b,*** *Amphibolitic intercalations in the lower part of the section;* ***c,*** *Grey foliated limestones;* ***d,*** *Ampelite interval;* ***e*** *and* ***f,*** *Strongly folded greywackes of the upper part of the Pardalà section.*

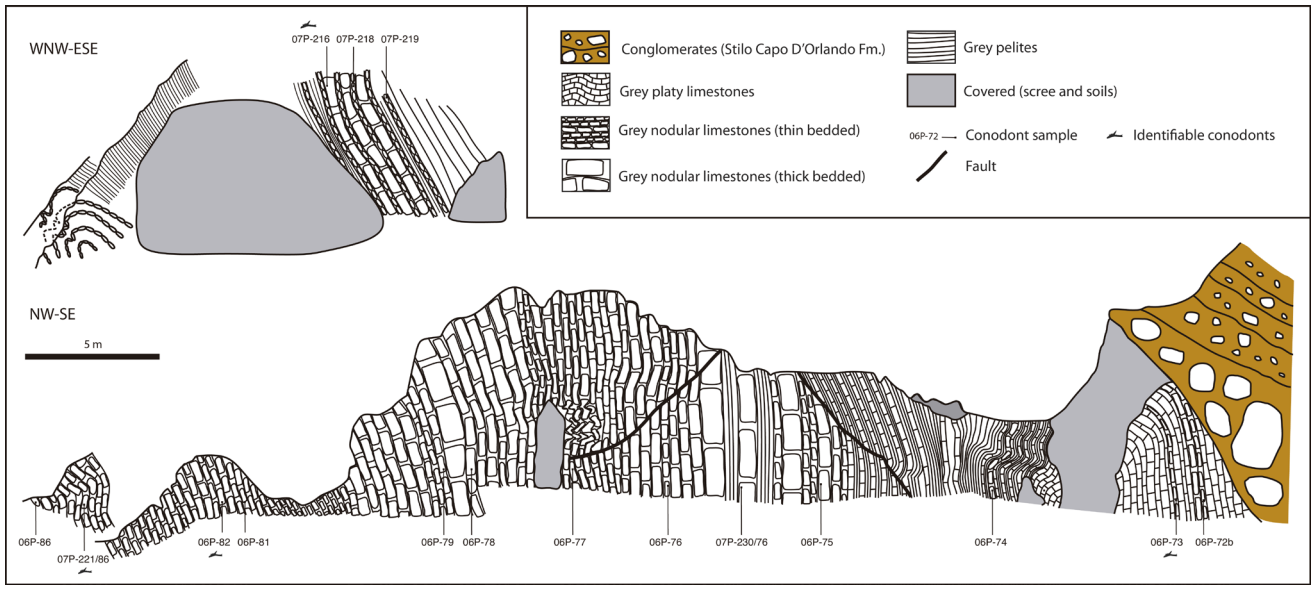


Figure III.1.5. Geological cross section along the Fiumara Assi section with location of the studied samples.

interval (partially covered at its base), formed by an alternation of thin-bedded calcareous slates, with aligned calcareous nodules, and centimetre- to decimetre-thick dark-grey limestone beds, locally with nodular structure (Fig. III.1.6.b). Upwards, the succession gradually becomes more calcareous, and forms a thickening-upwards, thin- to medium-bedded succession of limestones with progressively thinner intercalations of black pelites. This lithologic interval is followed by several tens of metres of limestones with very rare, centimetre- thick, grey to yellowish metapelite intercalations. Limestones are predominantly nodular (Fig. III.1.6.c) and contain abundant orthoceratids (Fig. III.1.6.e and g) and some isolated bivalves and corals (Fig. III.1.6.f). In the upper part of this interval, uppermost Pragian-Emsian conodonts have been found (Lower Devonian; see section III.1.4 for details).

The upper part of the succession, nearly 75 m thick and partially covered, consists mainly of grey to dark gray metapelites alternating with two nodular limestone intervals. Those of the lowermost interval have yielded Middle-Upper Devonian conodonts (see section III.1.4 for biostratigraphic details), gradually change to the under- and overlying metapelites, and are medium- to thin- bedded. The limestones of the uppermost interval are strongly affected by folds with metre to decimetre wavelength, and show a sharp cleavage and a strongly sheared contact with the metapelites (Fig. III.1.6.d). Structural and biostratigraphic data demonstrate that the Assi succession is overturned (Navas-Parejo *et al*., 2009a, see appendix I and section III.1.4 for biostratigraphic details).

San Giovanni 1 and 2 sections

Along an unnamed road from the Ponte Vina to the San Giovanni monastery, very weathered grey pelites crop out badly between soils. A few tens of metres after the road curve for the San Giovanni monastery, the San Giovanni 1 succession becames more carbonatic (4 in Fig. III.1.1.B; Fig. III.1.7) and contains platy marly limestone intercalations (Fig. III.1.9.a and b) bearing dacryoconarids and Middle Devonian conodonts (see section III.1.4 for biostratigraphic details). The succession is strongly folded and foliated, and shows cleavage related to axial planes of folds. Because of this deformation it is hard to establish the stratigraphic thickness of the section, being in any case lower than 25 m. The last ten of metres in this section correspond to a reversed succession of thin-bedded light grey limestones (Fig. III.1.9.c) where Frasnian conodonts have been found (see section III.1.4 for biostratigraphic details).

Reddish nodular limestones crop out in a few hundred metres lens along the same unnamed road (3 in Fig. III.1.1.B; Fig. III.1.8), a few tens of metres from the San Giovanni 1 section. They form an antiformal syncline (Fig. III.1.9.d) since conodonts found are progressively younger to the center of the fold (see section III.1.4 for biostratigraphic details) and, consequently, the stratigraphic succession of this San Giovanni 2 section is upside down. Therefore this antiformal syncline is probably a portion in the reversed limb of a major anticline structure.

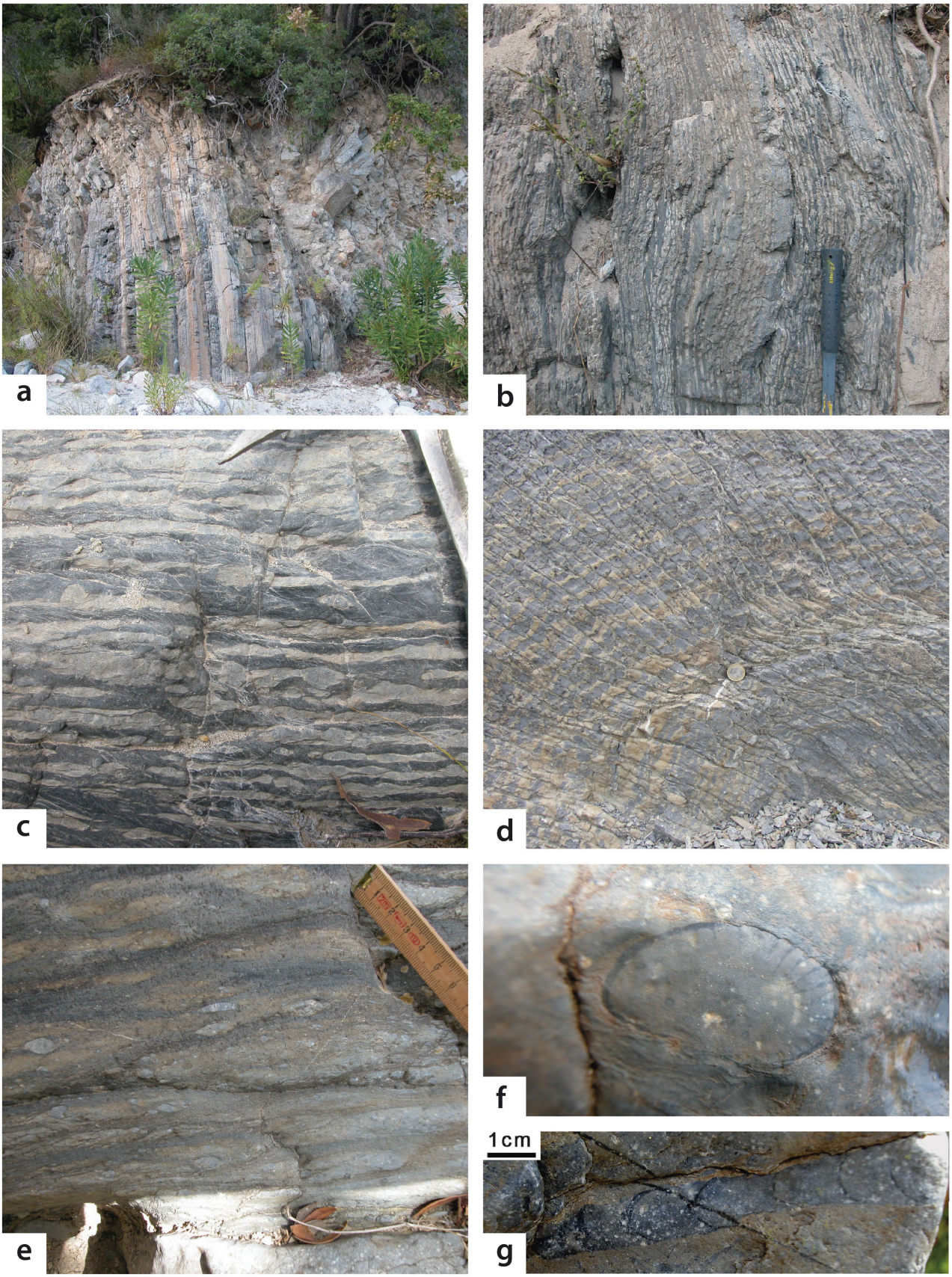


Figure III.1.6. Assi section. ***a,*** *Unconformity of the reversed stratigraphic lower part of the reversed Fiumara Assi succession (left) below the Stilo-Capo d'Orlando Fm (right);* ***b,*** *Thin-bedded nodular limestones;* ***c,*** *Nodular limestones of the upper part of the more carbonate interval;* ***d,*** *Strongly foliated thin bedded nodular limestones of the upper part of the section;* ***e,*** *Flattened orthoceratid transversal sections;* ***f,*** *Flattened isolated coral;* ***g,*** *Orthoceratid longitudinal section.*

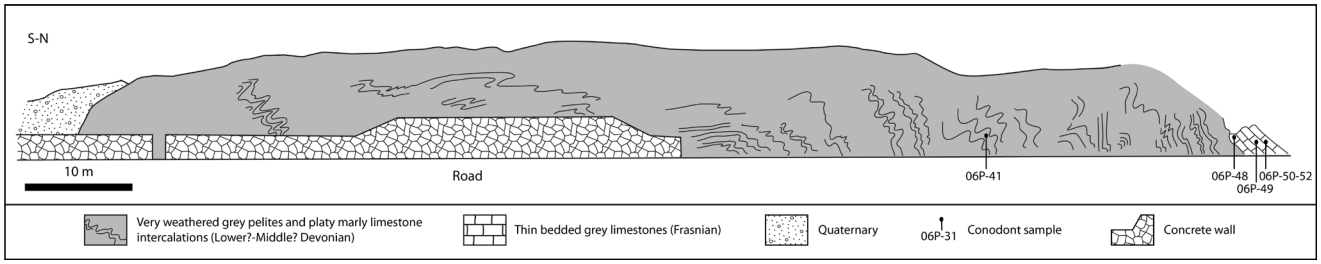


Figure III.1.7. San Giovanni 1 section.

These nodular limestones (*griotte* facies; figs. III.1.9.e and f) correspond to those dated as Famennian by Majesté-Mejoulas *et al*. (1984) and Bouillin *et al*. (1987).

Ponte Vina section

Along the road from Bivongi to Monasterace (SP 95), on a bend near the Vina bridge, downwater of the confluence of the Torrente Pardalà and the Fiumara Stilaro (6 in Fig. III.1.1.B; Fig. III.1.10), the same section previously dated by de Capoa-Bonardi (1970) has also been studied but including its lowermost part, where the Upper Kellwasser event has been found (see section III.1.4 for biostratigraphic details, and Navas-Parejo *et al.*, 2009b, appendix I). The 8-m-thick succession is made up of thin- to medium-bedded grey limestones with thin pelitic partings. The bedding of the slightly deformed succession is regular and shows no evidence of tectonic omission. The base of the section emerges from the alluvial sediments of the Fiumara Stilaro and starts with a 50-cm-thick interval composed of thin-bedded grey limestones, followed by a 20-cm-thick, black clayey horizon with millimetre- to centimetre-thick intercalations of grey limestones (Fig. III.1.12.a). This interval has released upper Frasnian conodonts (see chapter III.1.4 for biostratigraphic details). The thin pelitic interval is overlain by 20-cm-thick grey limestones, where Famennian conodonts have been found (see section III.1.4 for biostratigraphic details). Therefore, the black pelitic interval is correlated to the Upper Kellwasser event. The uppermost part of the section is nearly 7 m thick, and it is made up exclusively of dark-grey to bluish, thin- to medium-bedded limestones, which are slightly nodular and bioturbated in places (Fig. III.1.12.b and c).

Stilaro section

A few hundreds of metres to the NE of the Ponte Vina, also along the Fiumara Stilaro, a limestone lens crop out with a thickness of several tens of metres (6 in Fig. III.1.1; Fig. III.1.11). The succession, strongly dipping to the SE or nearly vertical, is made up of thin to medium bedded grey limestones, sightly nodular to the base of the succession, which also includes several dark pelitic intervals (figs. III.1.12 e and f). Famennian conodonts have been found in one sample taken from the slightly nodular limestones. Further sampling and studies are needed in order to confirm the record of the Kellwasser events also in this section.

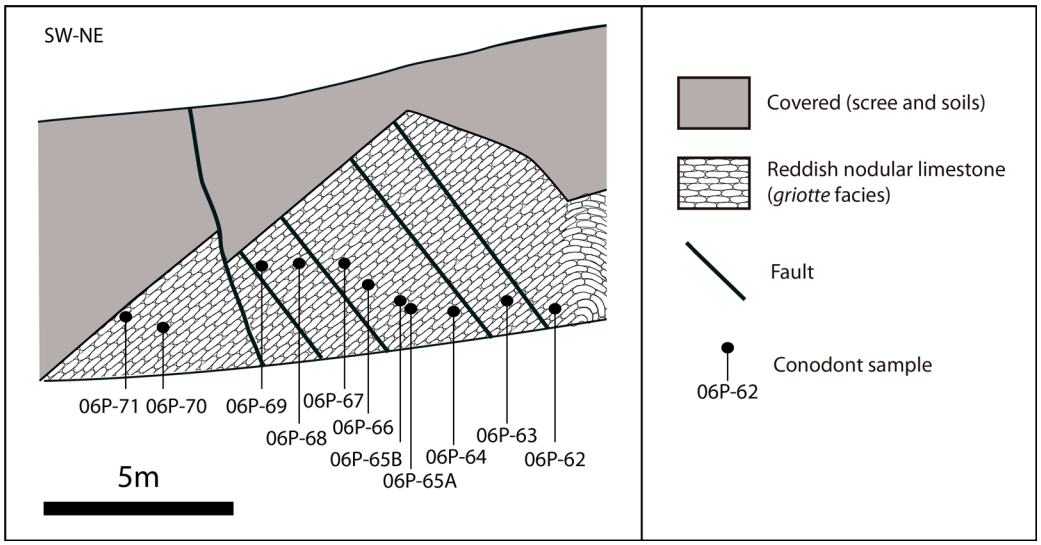


Figure III.1.8. San Giovanni 2 section.

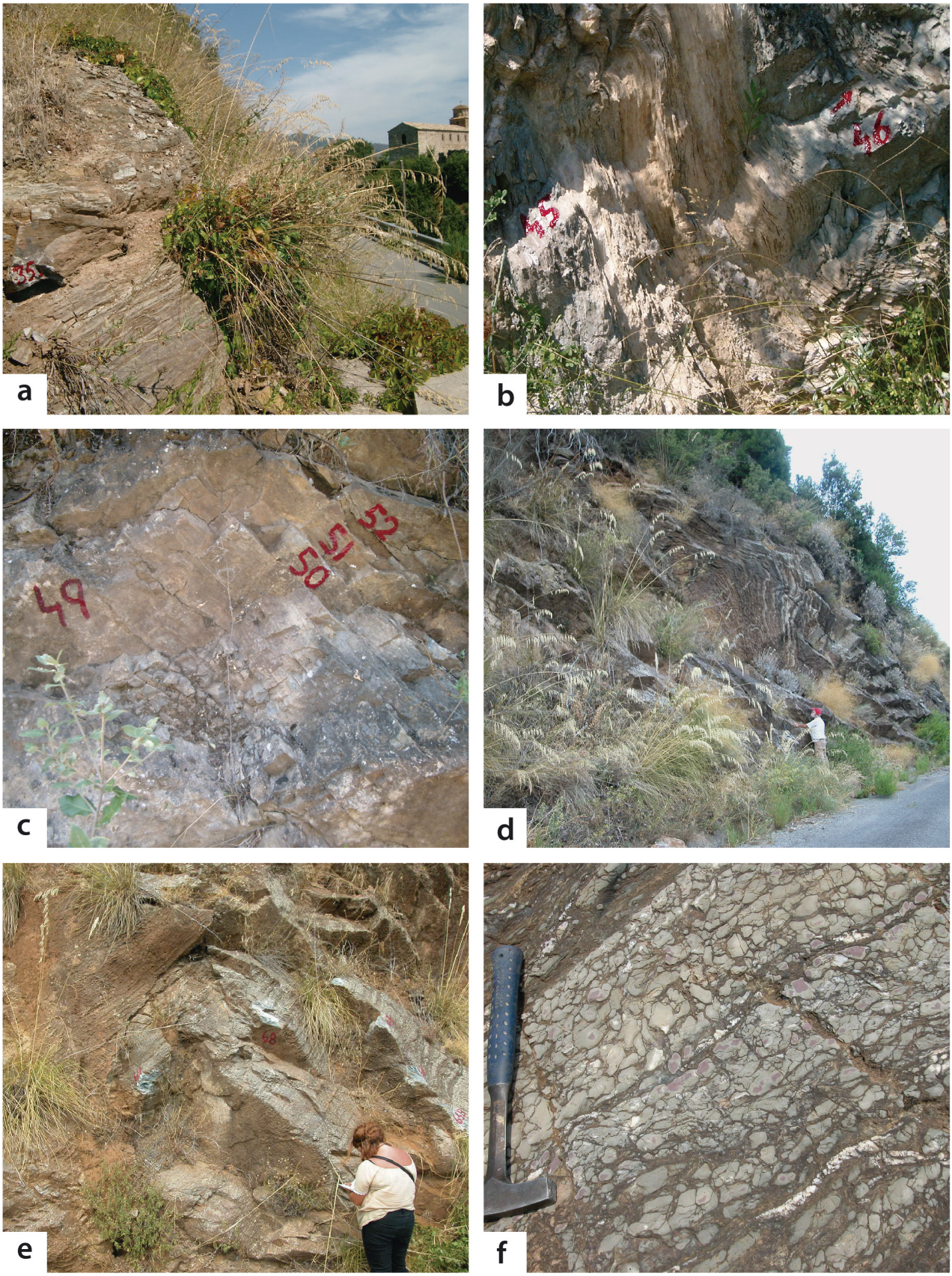


Figure III.1.9. San Giovanni 1 and 2 sections. ***a,*** *Very thin-bedded limestones and slaty alternations and the San Giovanni monastery in the background;* ***b,*** *Upper part of the more pelitic succession;* ***c,*** *Thin bedded limestones that yielded Frasnian conodonts (stratigraphically reversed beds);* ***d,*** *Fold in the San Giovanni 2 section;* ***e,*** *Very thin bedded reddish nodular limestones that released Famennian conodonts (reversed succession);* ***f,*** *Detail of e.*

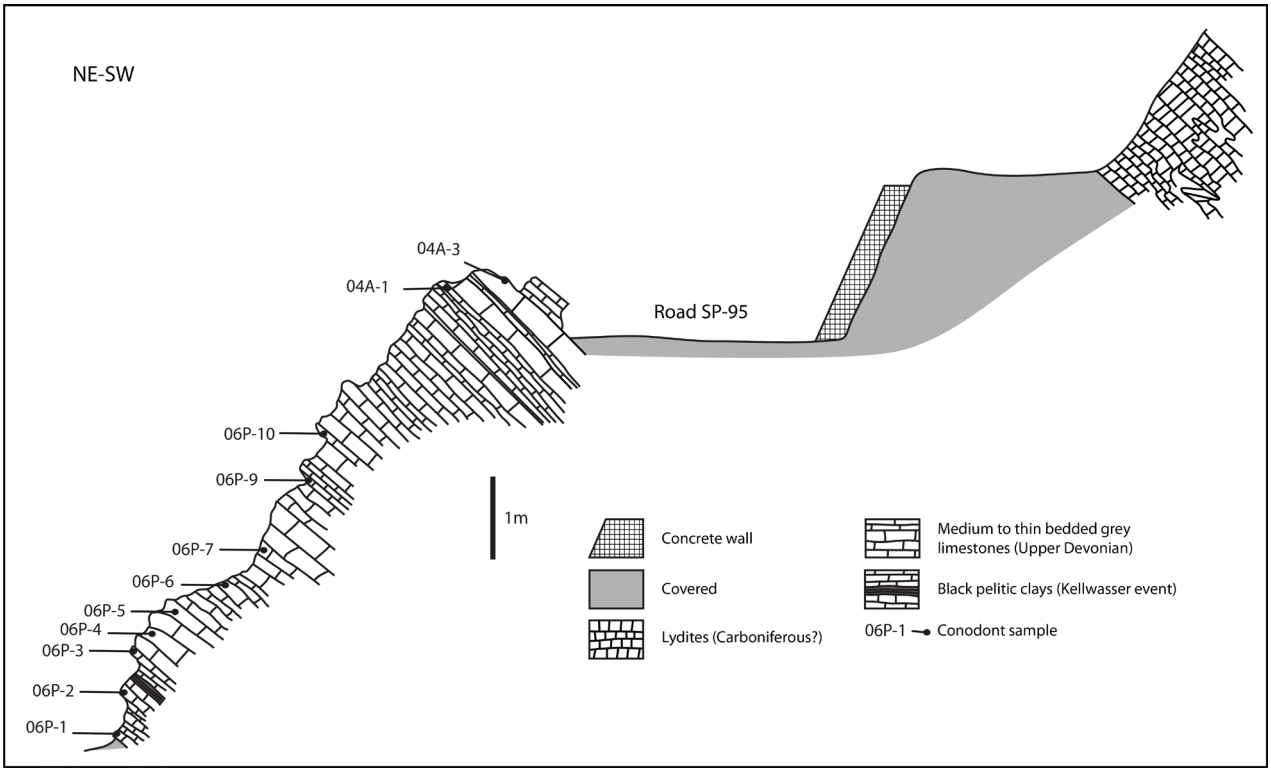


Figure III.1.10. Ponte Vina section.

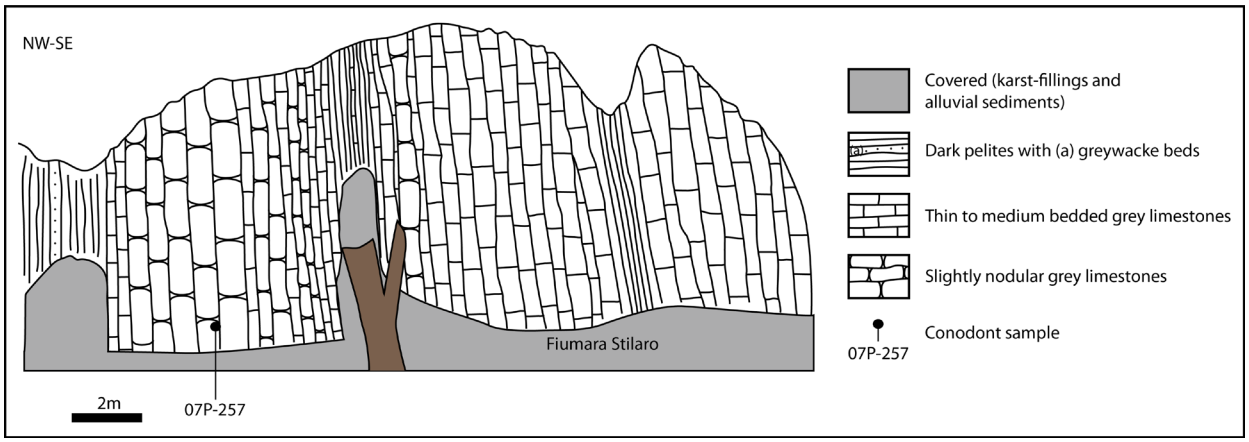


Figure III.1.11. Stilaro section.

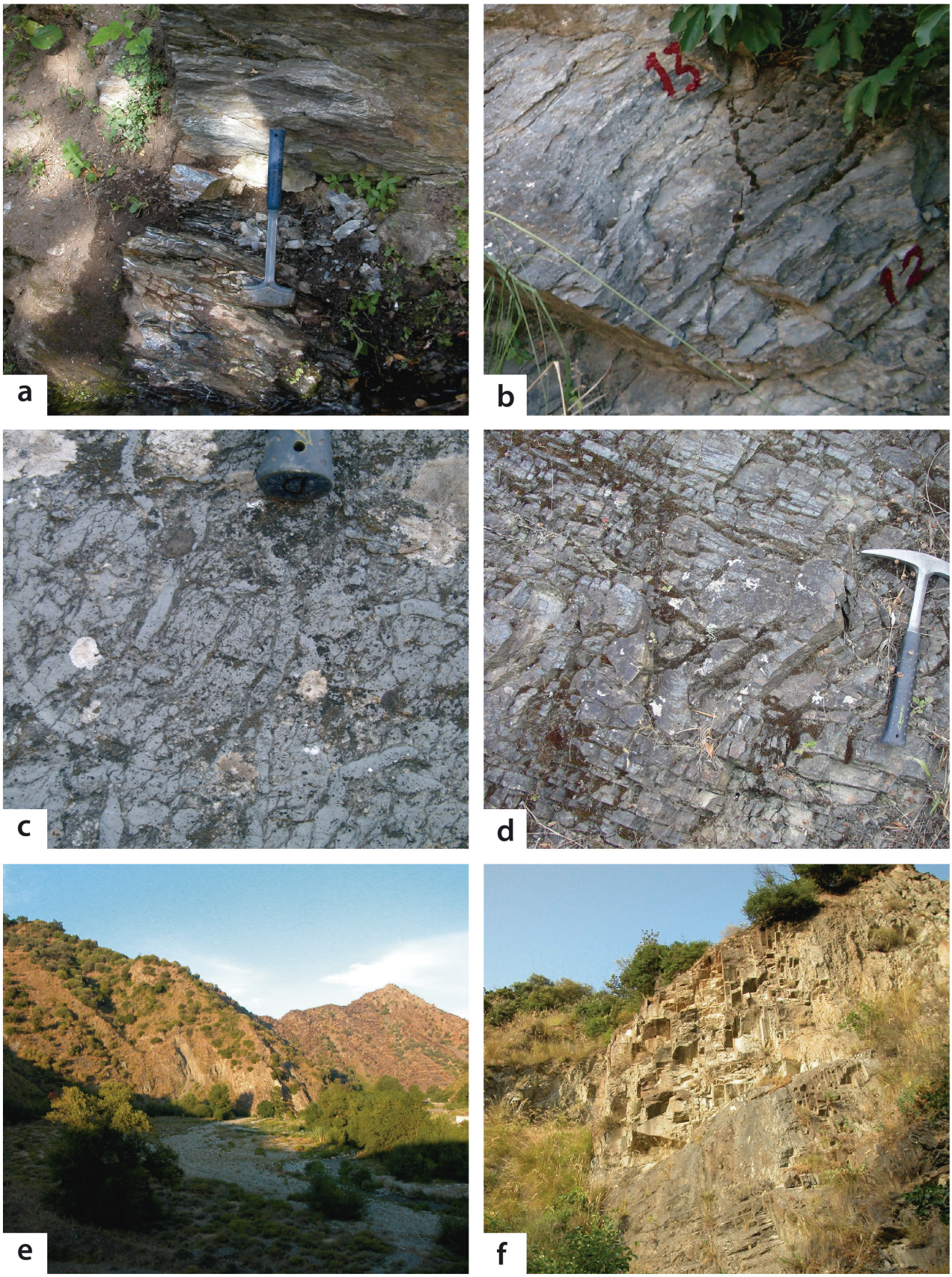


Figure III.1.12. Ponte Vina (a-d) and Stilaro (e-f) sections. ***a,*** *Thin bedded limestones that yielded Upper Devonian conodonts. The black pelitic clay interval (it coincides with the middle part of the hammer) corresponds to the Upper Kellwasser event;* ***b,*** *Medium bedded limestones which yielded Famenian conodonts;* ***c,*** *Bioturbated bedding surface in the upper part of the section;* ***d,*** *Folded lydites (black cherts) presumably of Early Carboniferous age;* ***e,*** *Panorama of the Fiumara Stilaro from the Ponte Vina to the NE, with the Stilaro limestones (left) dipping to the SE;* ***f,*** *Stilaro limestone interval where Famennian conodonts have been found.*

III.1.4. Conodont biostratigraphy

A total of a hundred and sixteen samples were taken in the six sections described above and fifty one of them released conodonts. Several biostratigraphic intervals belonging to the Devonian have been characterized in the Stilo Unit of the Bivongi-Stilo area.

Assi section

Scarce and poorly preserved conodont elements have been found in eighteen of forty four limestone samples studied in the Assi section (Fig. III.1.5). Most of these can be classified only at the generic level because of strong deformation. The most abundant elements belong to *Pseudooneotodus beckmanni* (Bischoff and Sannemann, 1958) and *Belodella* sp. Ethington, 1959.

The oldest association has been found in sample 06P-73. It is formed by *Icriodus* cf. *steinachensis* Al Rawi, 1977 and *Pelekysgnathus serratus* Jentzsch, 1962, which occurred in the *delta-sulcatus* zones (Lochkovian–Lower Pragian; Lower Devonian). Pa specimen figured in Fig. III.1.13.1 of *I.* cf. *steinachensis* is badly preserved but it shows a spindle-shaped outline very similar to that of the eta morphotype of the species (Klapper and Johnson, 1980). The angle of its lateral process with the main axis of the spindle is slightly less than 90º and denticles are rounded.

The different subspecies of *Pelekysgnathus* belonging to the *serratus* group are distinguished mainly on the basis of the size and position of the posterior denticles of the blade (Ziegler, 1991), which are not preserved in the specimens found (Fig. III.1.13.5). Consequently, there are no

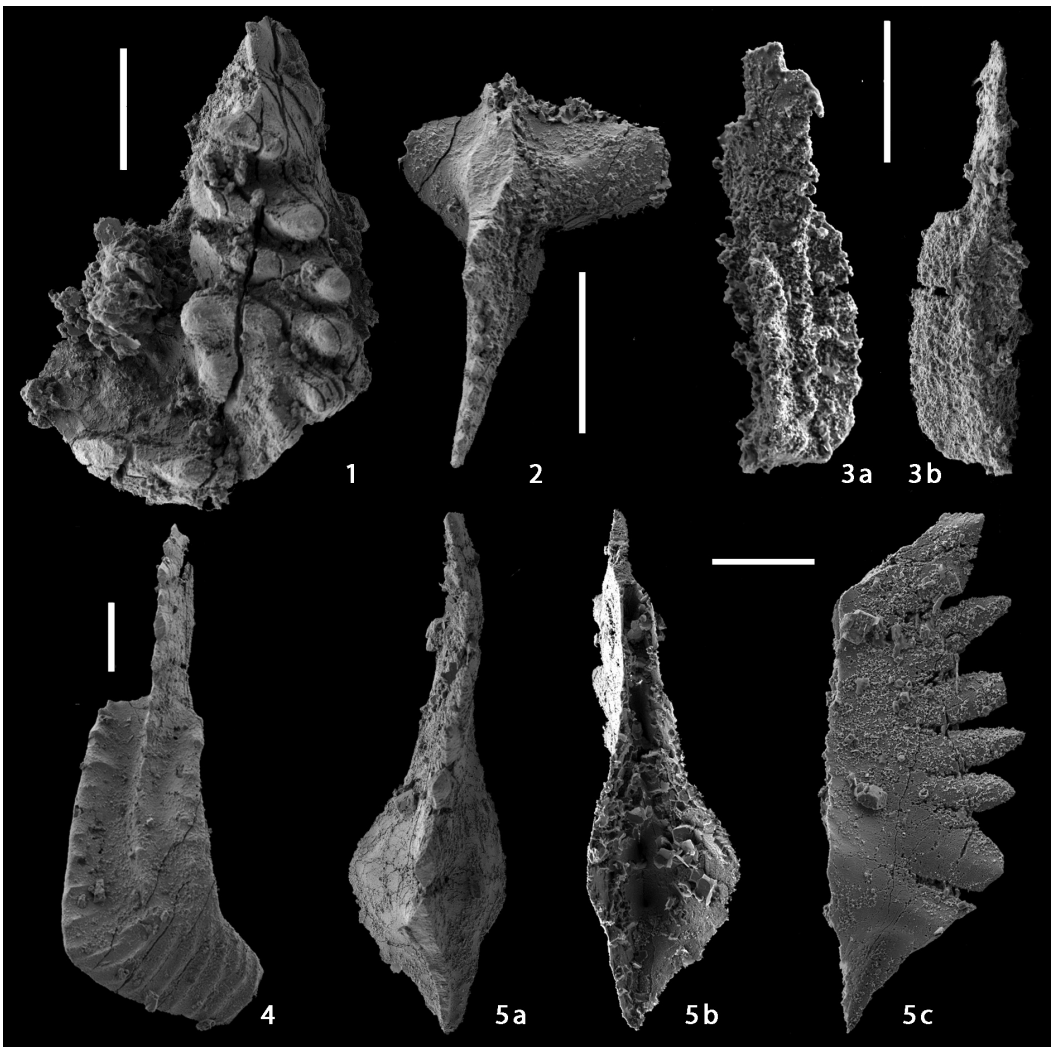


Figure III.1.13. Devonian conodonts from the Assi section. Scale bars are 200 µm. All are Pa elements. ***1****, Icriodus cf. steinachensis* eta morphotype*; 06P-73-4.* ***2****, 'Ozarkodina'* cf. *miae; 07P-221/86- 1.* ***3****, Polygnathus* cf. *kitabicus; 3a, upper view; 3b, lower view; 06P-82-3.* ***4****, Polygnathus linguiformis linguiformis; 07P-216-1;* ***5****, Pelekysgnathus serratus; 06P-73-1; 5a, upper view; 5b, lower view; 5c, lateral view.*

indisputable criteria to assign them to any of the subspecies defined within this group.

Geometrically below the former, younger conodont associations have been found. First, several Pa elements of *Polygnathus* cf. *kitabicus* Yolkin, Weddige, Izokh and Erina, 1994 appear in sample 06P-82. Despite its bad preservation, the illustrated specimen shows a large basal cavity and its carina does not reach the posterior end (Fig. III.1.13.3). This species ranges from the *kitabicus* Zone to the lowermost part of the *dehiscens* Zone, indicating the lowermost part the lower Emsian (Lower Devonian).

One Pa element of *'Ozarkodina'* cf. *miae* (Bultynck, 1971), which belongs to the *dehiscens-inversus* zones, has been found several metres upwards, in sample 07P-221/86. Although its anterior half is broken, typical asymmetrical oval lobes are visible (Fig. III.1.13.2).

The youngest conodont association in the Assi section has been identified in sample 07P-216, located at the geometric base of the outcrop. This association is characterised by *Polygnathus linguiformis linguiformis* Hinde, 1879, a species occurring from the Eifelian to the lower Frasnian.

The vertical distribution of the conodonts recorded in this section indicates that the Assi stratigraphic succession has been tectonically reversed.

San Giovanni 1 section

In the San Giovanni 1 section (Fig. III.1.7), seven of twenty seven samples released identifiable Devonian conodonts.

The lower one corresponds to sample 06P-41, which has yielded one Pa element of *Polygnathus*. In spite of its bad conservation, it could be assigned to *Po. serotinus* Telford, 1975 although with some reservations. Specimen figured in Fig. III.1.14.1 shows the characteristic shape of the platform for this species, and the typical wide and deep adcarinal troughs (Ziegler, 1977). Unfortunatelly, the basal cavity is very bad preserved although it seems to be inverted. Hence, the presence of this species preliminarly indicates a late Lower Devonian-early Middle Devonian age for this bed, even though further sampling and studies are needed.

The following bed that yielded conodonts coincides with sample 06P-48. The association includes *Palmatolepis praetriangularis,* Ziegler and Sandberg, 1988, *Palmatolepis triangularis* Sannemann, 1955a, *Palmatolepis delicatula delicatula* Branson and Mehl, 1934, *Icriodus alternatus alternatus* Branson and Mehl, 1934 and *Icriodus alternatus helmsi* Sandberg and Dressen, 1984, which belong to the Middle *triangularis* Zone (lower Famennian; Upper Devonian).

Typical features of *Palmatolepis praetriangularis* Pa elements are a sigmoidal carina, nodules in the inner platform and a flat or downward flexed posterior platform (Sandberg *et al.*, 1988) (Fig. III.1.14.2). This species differs from *Palmatolepis triangularis*, also found in this sample, by the posterior platform that is flexed upwards in this latter (Over and Rhodes, 2000) (Fig. III.1.14.3). *Palmatolepis delicatula delicatula* is characterised by Pa elements with a rounded outer lobe and a small but strong platform (Savage *et al.*, 2006). The outer lobe is not individualized and the whole platform has a quite triangular outline (Fig. III.1.14.4). In Pa elements of *Icriodus alternatus alternatus*, denticles of the medium row are extreme compressed laterally and alternate with the ones of the lateral rows (Fig. III.1.14.5). *Icriodus alternatus helmsi* Pa elements differ in the cusp that is aligned with a lateral row instead of the median one (Savage, 1992) (Fig. III.1.14.6).

Stratigraphically above, samples 06P-49, 06P-50 and 06P-51 released Frasnian conodonts. Sample 06P-49 yielded *Palmatolepis* cf. *rhenana brevis* Ziegler and Sandberg, 1990 and *Ancyrodella curvata* (Branson and Mehl, 1934). This fauna association belongs to the *rhenana* Zone (late Frasnian; Upper Devonian).

Specimen 7 in Fig. III.1.14 is assigned to *Palmatolepis rhenana brevis* with reservations, since it shows the typical outline of the platform, a sigmoidal carina and a short and rounded posterior end, but the external lobe is less differentiated (Ziegler and Sandberg, 1990). *Ancyrodella curvata* is characterised by Pa elements with a well-developed forth external lobe with carina and keel (Klapper, 1989) (Fig. III.1.14.8).

Sample 06P-50 released *Palmatolepis* cf. *jamieae* Ziegler and Sandberg, 1990, *Palmatolepis* cf. *delicatula* Branson and Mehl, 1934, *Ancyrognathus triangularis* (Youngquist, 1945), *Ancyrodella curvata* (Branson and Mehl, 1934) and *Ancyrodella nodosa* (Ulrich and Bassler, 1926), which belong to the Upper *rhenana* Zone (upper Frasnian; Upper

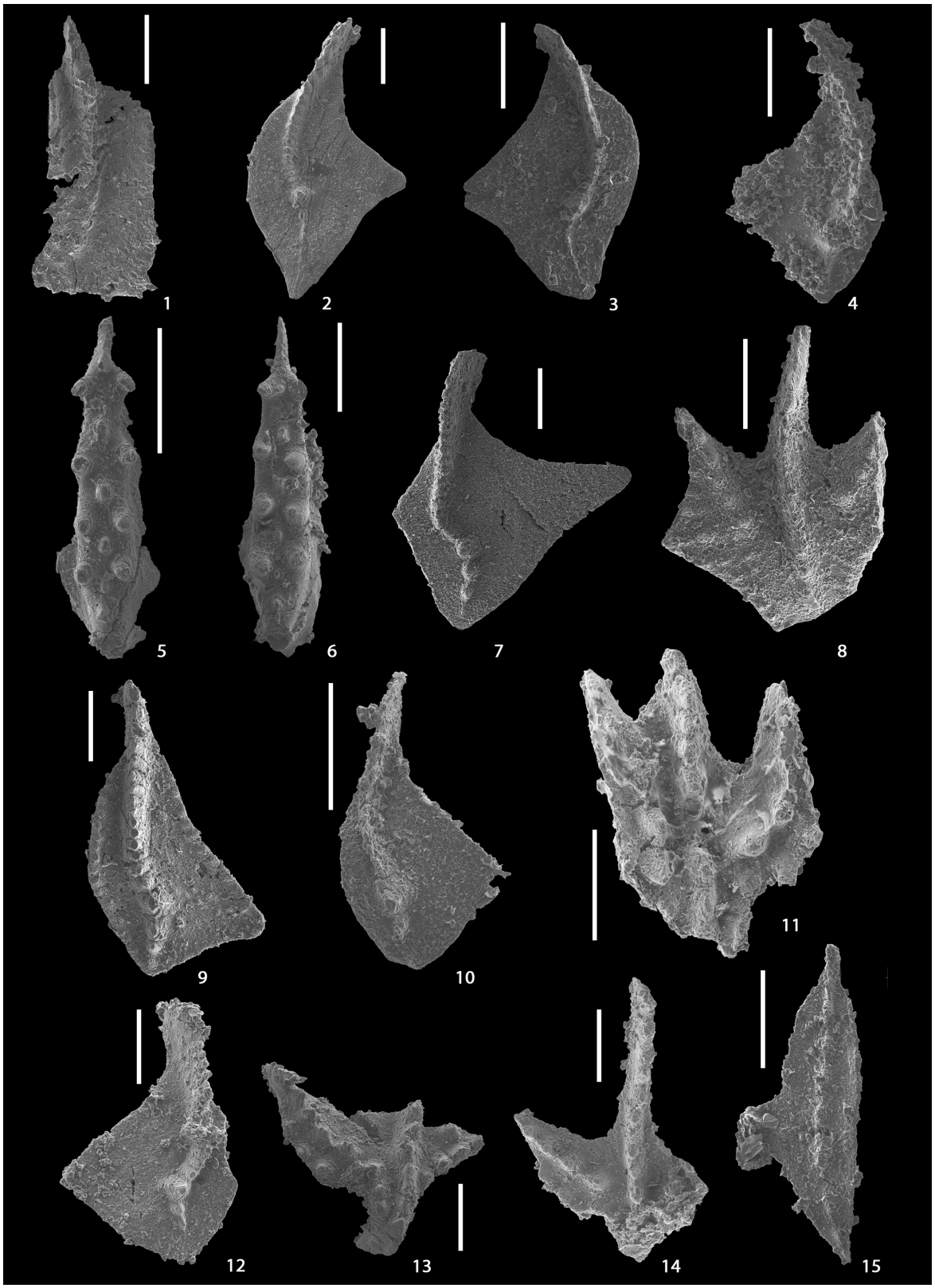


Figure III.1.14. Devonian conodonts of the San Giovanni 1 section. Scale bars are 200 µm. All are Pa elements in oral view. ***1,*** *Polygnathus* cf. *serotinus; 06P-41-1.* ***2,*** *Palmatolepis praetriangularis; 06P-48-1.* ***3,*** *Palmatolepis triangularis; 06P-48-3.* ***4,*** *Palmatolepis delicatula delicatula; 06P-48-4.* ***5,*** *Icriodus alternatus alternatus; 06P-48-6.* ***6,*** *Icriodus alternatus helmsi; 06P-48-7.* ***7,*** *Palmatolepis* cf. *rhenana brevis; 06P-49-1.* ***8,*** *Ancyrodella curvata; 06P-49-2.* ***9,*** *Palmatolepis* cf. *jamieae*; *06P-50-1.* ***10,*** *Palmatolepis* cf. *delicatula; 06P-50-3.* ***11,*** *Ancyrodella curvata; 06P-50-6.* ***12,*** *Palmatolepis* cf. *delicatula; 06P-51-1.* ***13,*** *Ancyrodella ioides; 06P-51-4.* ***14,*** *Ancyrodella* cf. *nodosa; 06P-51-2.* ***15,*** *Ancyrognathus triangularis; 06P-50-4.*

Devonian).

The Pa element figured as *Palmatolepis* cf. *jamieae* shows characteristic features of this species, such as very developed anterior denticles, a protuberance in the inner posterior platform and nodules or ridges all over the edge (Ziegler and Sandberg, 1990), but the position of that protuberance is more posterior than in the nominative species (Fig. III.1.14.9). *Ancyrognathus triangularis* is characterised by Pa elements that show a obtuse angle between posterior and secondary carinae; posterior carina is directed to inner side (Klapper, 1990) (Fig. III.1.14.11). In spite of being partially broken, the Pa element found of *Ancyrodella nodosa* (Fig. III.1.14.13) presents three lobes: the posterior one is short and the lateral ones are narrow and anteriorly directed. The upper surface ornamentation consists of nodules rows in the edge of the lobes (Over, 2007). *Ancyrodella curvata* is identified by Pa elements with a well-developed forth lobe, laterally (or posterolaterally) directed, which presents a carina and a keel (Olivieri, 1984). *Palmatolepis delicatula* belongs to the Famennian *triangularis* Zone, although Ziegler and Sandberg (1990) found Palmatolepis Pa elements similar to *Pa*. *delicatula* in the uppermost Frasnian, like seems to happen also in this section (Fig. III.1.14.10).

The uppermost sample that yielded conodonts in the San Giovanni 1 section (sample 06P-51) includes *Palmatolepis* cf. *delicatula* Branson and Mehl 1934, *Ancyrodella* cf. *nodosa* (Ulrich and Bassler 1926) and *Ancyrodella ioides* Ziegler 1958. This conodont association belong to the Lower *rhenana* Zone. As occured in sample 06P-50, specimens very similar to *Pa. delicatula* have been found together with other Frasnian elements (Fig. III.1.14.14). As mentioned before, maybe it is necessary to expand the biostratigraphic distribution of this species into the uppermost Frasnian (Ziegler and Sandberg, 1990). The specimen of *Ancyrodella nodosa* found in this sample (Fig. III.1.14.15) has both lateral lobes partially broken, but presents all the other features that characterised this species, as is described above. *Ancyrodella ioides* differs from other *Ancyrodella* species by a reduced platform (Perri and Spalletta, 1980), feature that can be observed in the specimen here found although is partially broken (Fig. III.1.14.16).

San Giovanni 2 section

In the San Giovanni 2 section eleven samples were taken and all of them released identifiable conodonts. Several conodont biozones belonging to the lowermost Famennian have been identified (*crepida*, *rhomboidea* and *marginifera* zones). As mentioned in previous section, the biostratigraphic data indicates that the succession is reversed.

The oldest conodont associations in this section have been found in samples 06P-71, 06P-70 and 06P-69. All of them indicate an age that ranges from the upper *crepida* to the Uppermost *crepida* zones.

Sample 06P-71 has yielded *Palmatolepis quadrantinodosalobata* Sannemann, 1955a, *Palmatolepis glabra glabra* Ulrich and Bassler, 1926, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis minuta subgracilis* Bischoff, 1956, *Palmatolepis minuta wolskae* Szulczewski, 1971, *Polygnathus glaber glaber* Ulrich and Bassler, 1926 and *Icriodus alternatus alternatus* Branson and Mehl, 1934.

Sample 06P-70 released *Palmatolepis crepida* Sannemann, 1955b, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis subperlobata* Branson and Mehl, 1934 and *Icriodus alternatus alternatus* Branson and Mehl, 1934.

Conodont association found in sample 06P-69 is very similar, and includes *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis quadrantinodosalobata* Sannemann, 1955a, *Palmatolepis subperlobata* Branson and Mehl, 1934 and *Icriodus alternatus alternatus* Branson and Mehl, 1934.

Geometrically below, sample 06P-68 has released a conodont association that ranges from the Upper *crepida* to the Lower *rhomboidea* zones. This association contains *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis minuta wolskae* Szulczewski, 1971, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis subperlobata* Branson and Mehl, 1934 and *Polygnathus* aff. *glaber* Ulrich and Bassler, 1926.

A similar biostratigraphic distribution, from the Uppermost *crepida* to the Lower *rhomboidea* zones is inferred from conodonts found in sample 06P-67, which yielded *Palmatolepis glabra pectinata* Ziegler, 1962a, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis minuta minuta* Branson

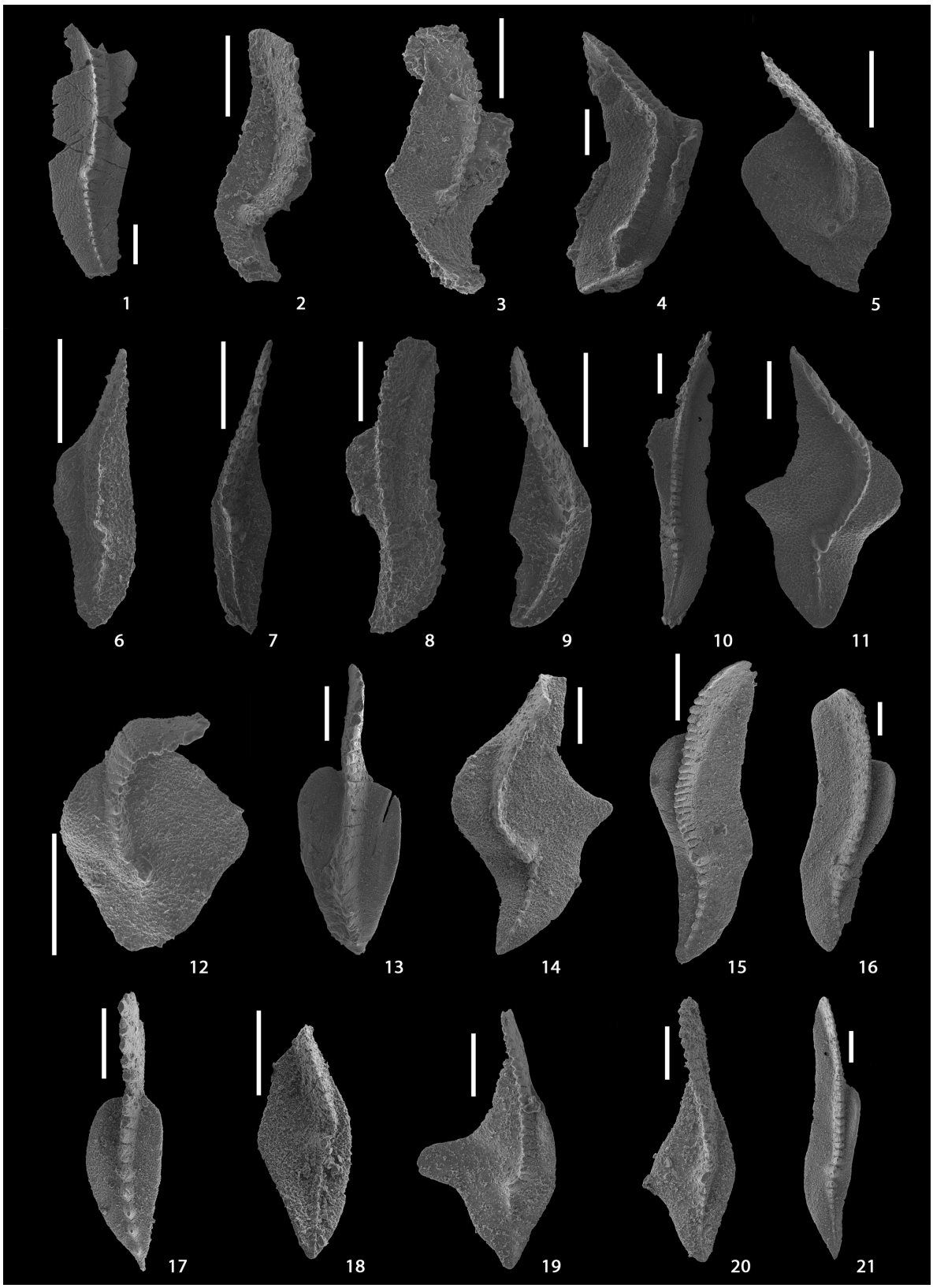


Figure III.1.15. Upper Devonian conodonts of the San Giovanni 2 section (plate 1). All are oral view of Pa elements. Scale bars are 200 µm. **1,** *Palmatolepis* cf. *falcata*; 06P-62-1. **2,** *Palmatolepis glabra pectinata*; 06P-62-2. **3,** *Palmatolepis glabra glabra*; 06P-62-3. **4,** *Palmatolepis glabra pectinata*; 06P-63-1. **5,** *Palmatolepis rhomboidea*; 06P-63-2. **6,** *Palmatolepis glabra prima*; 06P-63-3. **7,** *Palmatolepis minuta minuta*; 06P-63-4. **8,** *Palmatolepis glabra prima*; 06P-64-1. **9,** *Palmatolepis minuta minuta*; 06P-64-4. **10,** *Palmatolepis falcata*; 06P-65A-1. **11,** *Palmatolepis schindewolfi*; 06P-65A-2; **12,** *Palmatolepis rhomboidea*; 06P-65A-3. **13,** *Polygnathus glaber bilobatus*; 06P-65A-4. **14,** *Palmatolepis schindewolfi*; 06P-65B-1. **15,** *Palmatolepis glabra prima*; 06P-65B-3. **16,** *Palmatolepis glabra pectinata*; 06P-65B-4. **17,** *Polygnathus glaber glaber*; 06P-65B-7. **18,** *Palmatolepis minuta minuta*; 06P-65B-6. **19,** *Palmatolepis* cf. *subperlobata*; 06P-66-1. **20,** *Palmatolepis minuta minuta*; 06P-66-5. **21,** *Palmatolepis glabra pectinata*; 06P-66-3.

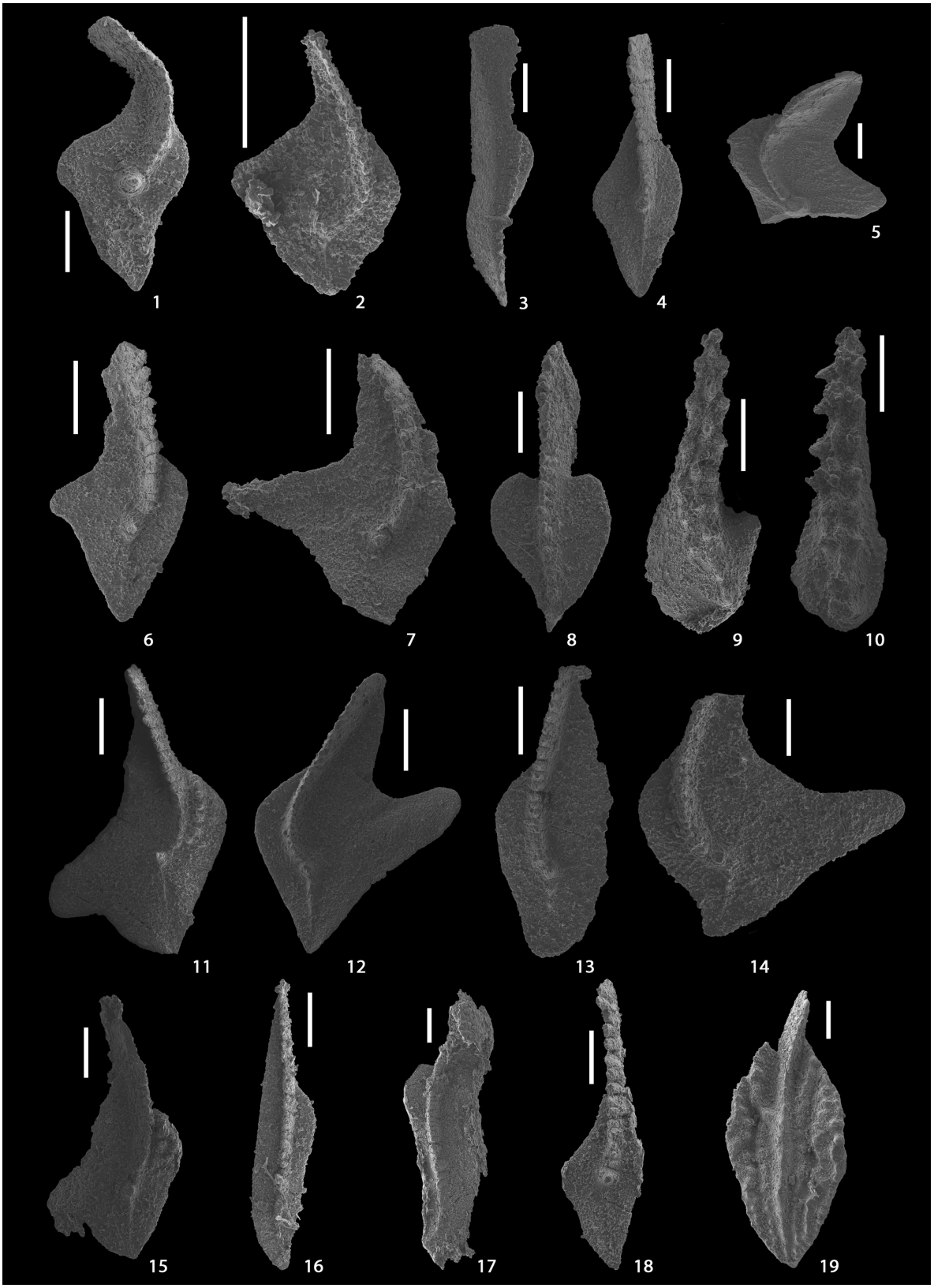


Figure III.1.16. Upper Devonian conodonts of the San Giovanni 2 (plate 2; photos 1-15) and Stilaro (photos 16-19) sections. All are oral view of Pa elements. Scale bars are 200 µm.

1, *Palmatolepis minuta wolskae*; 06P-66-6. **2,** *Palmatolepis rhomboidea*; 06P-66-7. **3,** *Palmatolepis glabra pectinata*; 06P-67-1. **4,** *Palmatolepis minuta minuta*; 06P-67-3. **5,** *Palmatolepis cf. subperlobata*; 06P-67-4. **6,** *Palmatolepis minuta wolskae*; 06P-68-1. **7,** *Palmatolepis subperlobata*; 06P-68-2. **8,** *Polygnathus* aff. *glaber*; 06P-68-3. **9,** *Icriodus alternatus alternatus*; 06P-69-1. **10,** *Icriodus alternatus alternatus*; 06P-70-2. **11,** *Palmatolepis quadrantinodosalobata*; 06P-69-3. **12,** *Palmatolepis subperlobata*; 06P-69-4. **13,** *Palmatolepis crepida*; 06P-70-1. **14,** *Palmatolepis subperlobata*; 06P-70-3. **15,** *Palmatolepis quadrantinodosalobata*; 06P-71-1. **16,** *Palmatolepis glabra prima*; 07P-257-3. **17,** *Palmatolepis glabra pectinata*; 07P-257-1. **18,** *Palmatolepis minuta minuta*; 07P-257-2. **19,** *Polygnathus* sp.; 07P-257-4.

and Mehl, 1934 and *Palmatolepis* cf. *subperlobata* Branson and Mehl, 1934.

Conodont association found in sample 06P-66 points at the Lower *rhomboidea* Zone, and contains *Palmatolepis rhomboidea* Sannemann, 1955a, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis minuta wolskae* Szulczewski, 1971, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis* cf. *subperlobata* Branson and Mehl, 1934, *Palmatolepis glabra pectinata* Ziegler, 1962a, *Palmatolepis schindewolfi* Müller, 1956 and *Polygnathus glaber glaber* Ulrich and Bassler, 1926.

Palmatolepis rhomboidea is characterised by Pa elements with a very small and oval platform (figs. III.1.15.5 and 13, and III.1.16.2). Blade and carina are sigmoidal, and there is a bulge or low parapet on the anterior inner platform, and the posterior end is flexed upwards (Ziegler, 1974). *Palmatolepis minuta wolskae* Pa elements show an outer lateral lobe, less differentiated than in *Pa. minuta loba*, and a prominent central node (Ziegler, 1977) (Fig. III.1.16.1 and 6). *Palmatolepis schindewolfi* includes large and elongate Pa elements with a lateral (outer) pointed lobe, moderately developed Sandberg and Ziegler, 1979) (Fig. III.1.15.11 and 14).

Sample 06P-65B released a conodont association that indicates an age that ranges from the Upper *rhomboidea* to the Upper *marginifera* zones. This association contains *Palmatolepis glabra pectinata* Ziegler, 1962a, *Palmatolepis schindewolfi* Müller, 1956, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis gracilis gracilis* Branson and Mehl, 1934 and *Polygnathus glaber glaber* Ulrich and Bassler, 1926.

Pa elements of *Polygnathus glaber glaber* are characterised by a lanceolate shape with deep troughs on both sides of the carina (Metzger, 1994) (Fig. III.1.15.17).

In sample 06P-65A the Lower *marginifera* Zone has been identified by the presence of *Palmatolepis rhomboidea* Sannemann, 1955a, *Palmatolepis falcata* (Helms, 1959b) and *Polygnathus glaber bilobatus* Ziegler, 1962b. The conodont association found in this sample also contains *Palmatolepis glabra glabra* Ulrich and Bassler, 1926, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis schindewolfi* Müller, 1956, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis gracilis gracilis* Branson and Mehl, 1934 and *Polygnathus glaber glaber* Ulrich and Bassler, 1926.

Polygnathus glaber bilobatus differs from other Polygnathus glaber species by the presence of a lobe in the anterior part of the platform, being the platform slightly assymetrical (Ziegler, 1975) (Fig. III.1.15.13).

Diversity of the fauna recovered from sample 06P-64 did not allow to be specific about the age interval. The conodont association in this sample points at an age that ranges from the Upper *crepida* to the Upper *marginifera* zones and contains *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis glabra glabra* Ulrich and Bassler, 1926 and *Palmatolepis minuta minuta* Branson and Mehl, 1934.

Palmatolepis glabra pectinata Ziegler, 1962a, *Palmatolepis rhomboidea* Sannemann, 1955a, *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis minuta minuta* Branson and Mehl, 1934, *Palmatolepis* cf. *gracilis* Branson and Mehl, 1934 were found in sample 06P-63. These species belong to the Upper *rhomboidea* Zone.

Close to the center of the antiform fold, sample 06P-62 has been taken. The conodont association found belong to the Lower *marginifera* Zone and contains *Palmatolepis* cf. *falcata* (Helms, 1959b), *Palmatolepis glabra pectinata* Ziegler, 1962a and *Palmatolepis glabra glabra* Ulrich and Bassler, 1926.

Palmatolepis glabra glabra is characterised by Pa elements with an elongated and narrow platform, without external lobe (Ziegler, 1977). In this *Pa. glabra* subspecies are included Pa elements with an external anterior end perpendicular to the blade (Fig. III.1.15.3).

Palmatolepis glabra pectinata includes elongated and slightly sigmoidal Pa elements that develop a parapet in the inner anterior platform (Ziegler, 1977). This parapet can be denticulate, although this feature has not taxonomic significance (figs. III.1.15.2, 4, 16, 21 and III.1.16.3, 17).

Palmatolepis falcata is characterised by Pa elements with a very narrow and elongate platform with a triangular parapet developed on the inner anterior platform that is bent upwards (Dzik, 1991) (Fig. III.1.15.1 and 10).

Ponte Vina section

In the Ponte Vina section, eleven of twenty

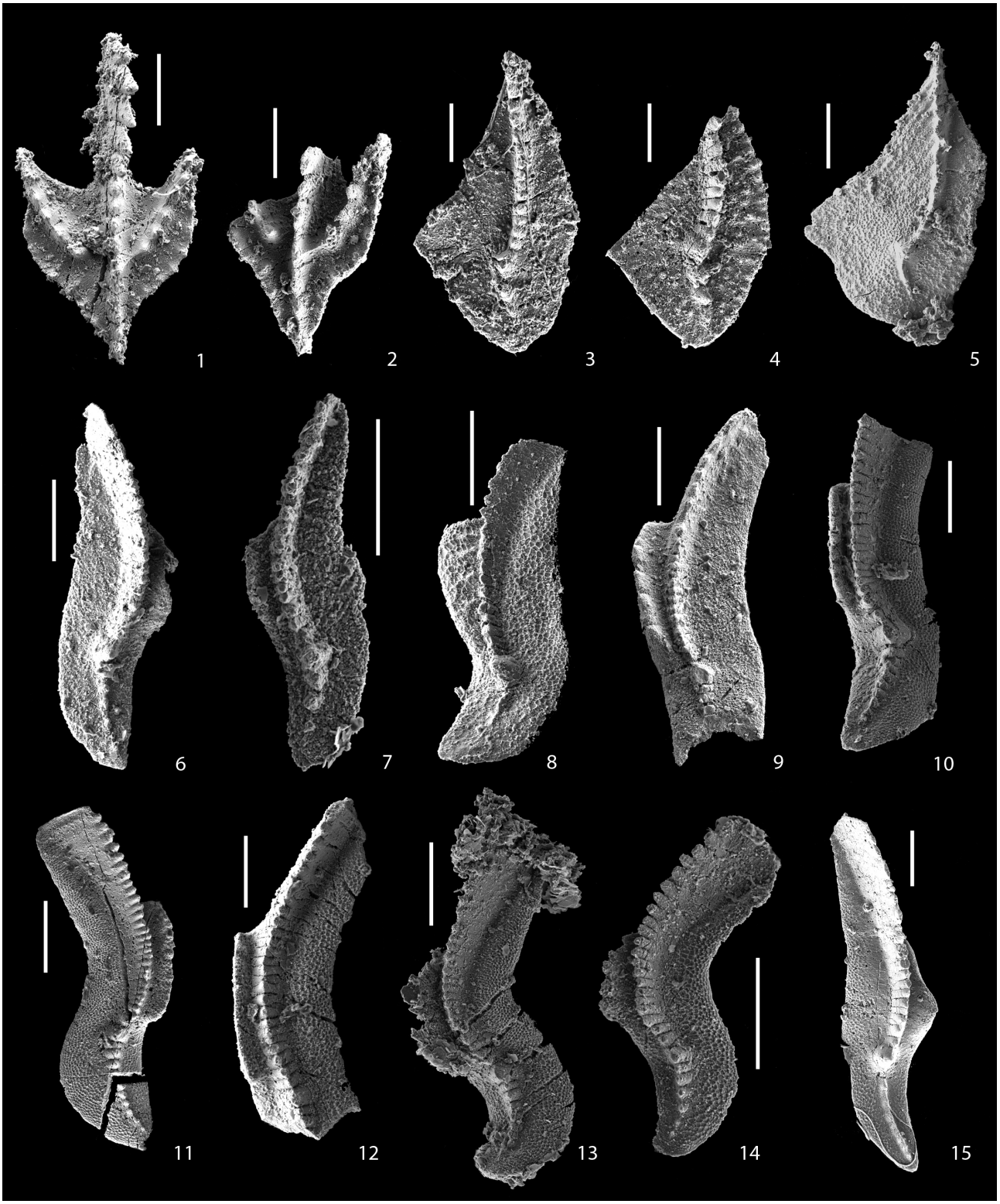


Figure III.1.17. Upper Devonian conodonts of the Ponte Vina section (plate 1). All are oral view of Pa elements. Scale bars are 200 µm.

***1,** Ancyrodella buckeyensis; 06P-1-1. **2,** Ancyrodella buckeyensis; 06P-1-9. **3,** Palmatolepis cf. uyenoi; 06P-1-5. **4,** Palmatolepis cf. uyenoi; 06P-1-6. **5,** Palmatolepis bogartensis; 06P- 2-1. **6,** Palmatolepis glabra prima; 06P-3-1. **7,** Palmatolepis glabra prima; 06P-6-14. **8,** Palmatolepis glabra prima; 06P-6-10. **9,** Palmatolepis glabra pectinata; 06P-3-4. **10,** Palmatolepis glabra pectinata; 06P-7-5. **11,** Palmatolepis glabra distorta; 06P-6-6. **12,** Palmatolepis glabra distorta; 06P-5- 10. **13,** Palmatolepis glabra distorta; 06P-6-13. **14,** Palmatolepis glabra distorta; 06P-7-4. **15,** Palmatolepis falcata; 06P-3-15.*

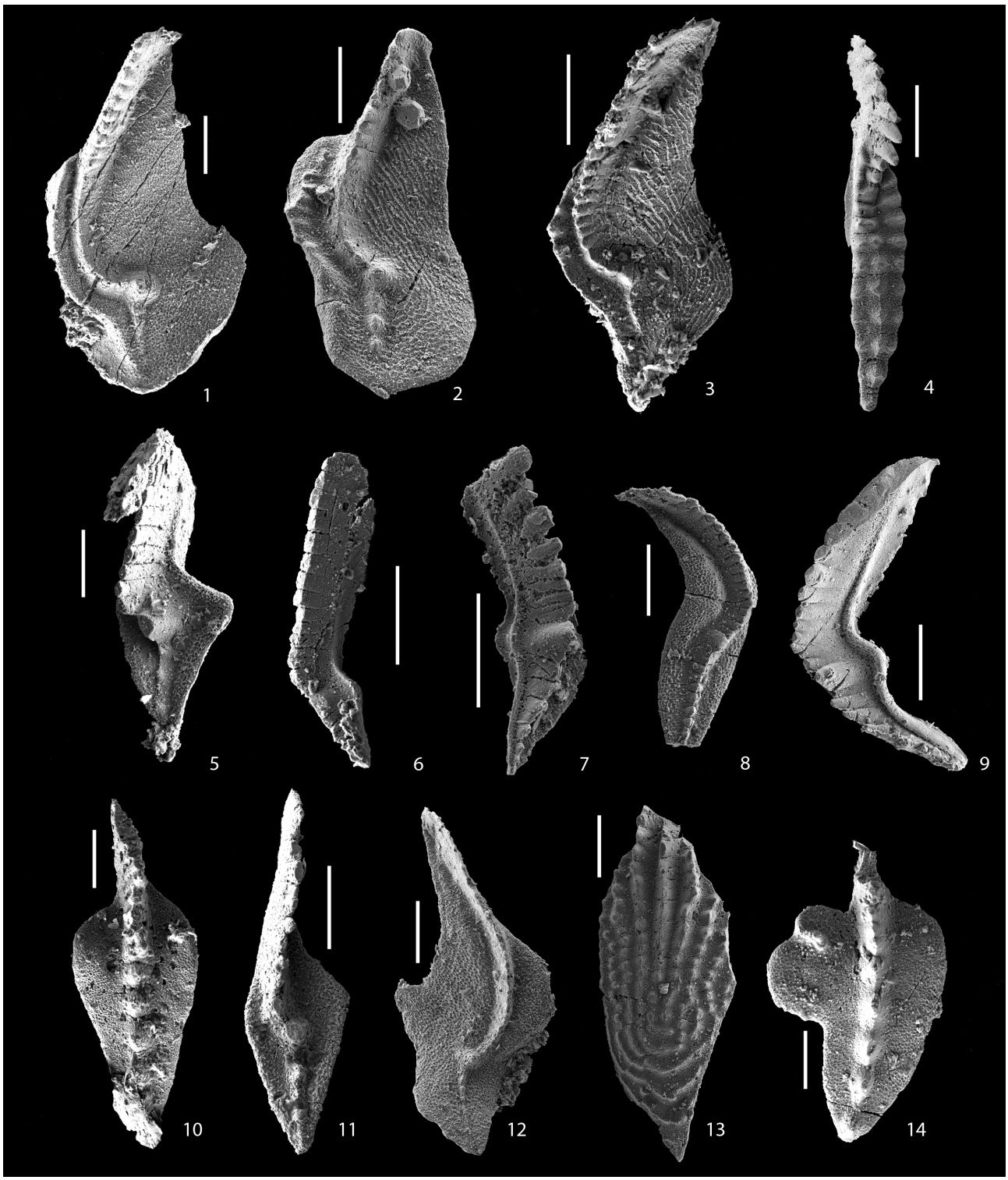


Figure III.1.18. Upper Devonian conodonts of the Ponte Vina section (plate 2). All are upper view of Pa elements. All scale bars are 200 µm. ***1,*** *Palmatolepis marginifera marginifera; 06P-3-8.* ***2,*** *Palmatolepis marginifera marginifera; 06P-4-6.* ***3,*** *Palmatolepis marginifera utahensis; 06P-6-3.* ***4,*** *Scaphignathus velifer velifer; 06P-9-3.* ***5,*** *Palmatolepis minuta schleizia; 06P-7-2.* ***6,*** *Palmatolepis gracilis gracilis; 06P-5-1.* ***7,*** *Palmatolepis gracilis gracilis; 06P-7-6.* ***8,*** *Palmatolepis gracilis manca; 04A-3-16.* ***9,*** *Palmatolepis gracilis sigmoidalis; 04A-3-23.* ***10,*** *Polygnathus glaber medius; 06P-3-7.* ***11,*** *Palmatolepis minuta minuta; 06P-6-7.* ***12,*** *Palmatolepis schindewolfi; 06P-3- 25.* ***13,*** *Polylophodonta cf. confluens; 06P-4-14.* ***14,*** *Polygnathus glaber bilobatus; 06P-5-5.*

five samples yielded conodonts. Those with more relevant biostratigraphic information are detailed below.

The lowermost bed of the section (sample 06P-1) contains *Ancyrodella buckeyensis* Stauffer, 1938 (Fig. III.1.17. 1 and 2), which ranges from the *punctata* Zone to the Upper *rhenana* Zone (equivalent to the MN 5 Zone and to the uppermost MN 12 Zone of Klapper, 1989). This sample also provided two broken Pa elements, which we assign tentatively to *Palmatolepis uyenoi* Klapper, 2007 (Fig. III.1.17. 3 and 4). The occurrence of this species identifies the MN 11–12 zones. The index conodont species of the MN 13 Zone, *Palmatolepis bogartensis* (Stauffer, 1938) (Fig. III.1.17.5), appears in sample 06P- 2. The black pelite interval can be seen just above this bed. All the conodont specimens recovered from the upper Frasnian horizons lying below the black pelite interval are pyritized whilst the samples taken from the limestone bed lying immediately above it are not. Unfortunately, the few conodont fragments released in the overlying bed cannot be determined and so do not allow accurate dating.

The first indisputable Famennian conodonts, obtained from the limestones overlying the black pelite bed, are included in sample 06P-3, located about 50 cm above sample 06P-2. These limestones have provided *Palmatolepis marginifera marginifera* Helms, 1959a (Fig. III.1.18.1). This is the defining subspecies for the base of the Lower *marginifera* Zone and ranges to the Uppermost *marginifera* Zone. Sample 06P-3 also contains *Polygnathus glaber medius* Helms and Wolska, 1967, which ranges from the Early to the Late *marginifera* zones. *Palmatolepis marginifera utahensis* Ziegler and Sandberg, 1984 (Fig. III.1.18.3), the index subspecies for the Upper *marginifera* Zone, first occurs in sample 06P-5. Sample 06P-7 contains, among other species, *Palmatolepis glabra pectinata* Ziegler, 1962a, whose upper range is the Upper *marginifera* Zone. The occurrence of *Scaphignathus velifer velifer* Beinert, 1971 in sample 06P-10 indicates the Uppermost *marginifera* Zone (Fig. III.1.18.4). We could not find any index taxa either for the *trachytera* or for the Lower *postera* Zone in sample 04A-1. Sample 04A-3, however, has provided *Palmatolepis gracilis manca* Helms, 1963 (Fig. III.1.18.8), which is the key subspecies for the base of the Late *postera* Zone. Consequently the interval between samples 06P-10 and 04A-3 cannot currently be definitively assigned to any zone.

Stilaro section

In the Stilaro section, three samples have been taken but only one released identifiable conodonts. The association of *Palmatolepis glabra prima* Ziegler and Huddle, 1969, *Palmatolepis glabra pectinata* Ziegler, 1962a and *Palmatolepis minuta minuta* Branson and Mehl, 1934 in sample 07P-257 indicates a range from the Upper *crepida* to the Middle *marginifera* (Famennian, Upper Devonian) for the medium-bedded nodular limestones in this section.

Section	Assi				San Giovanni 1				
Samples	06P-73	06P-82	07P-221/86	07P-216	06P-41	06P-48	06P-49	06P-50	06P-51
dissolved amount	2.450	0.725	0.700	0.700	0.200	0.325	0.325	0.275	0.200
Ancyrodella curvata							X	X	
Ancyrodella ioides									X
Ancyrodella nodosa								X	X
Ancyrognathus triangularis								X	
Icriodus alternatus alternatus						X	X		
Icriodus alternatus helmsi						X			
'Ozarkodina' cf. *miae*			X						
Palmatolepis delicatula delicatula						X		X (cf.)	X (cf.)
Palmatolepis jamiae								X (cf.)	
Palmatolepis praetriangularis						X			
Palmatolepis cf. *rhenana brevis*							X		
Palmatolepis triangularis						X			
Pelekysgnathus gr. *serratus*	X								
Polygnathus kitabicus		X							
Polygnathus linguiformis linguiformis				X					
Polygnathus serotinus					X				
Pseudooneotodus beckmanni	X	X		X					
Ancyrodella sp.							X	X	
Belodella sp.		X							
Palmatolepis sp./fragments						24	20	37	10
Polygnathus sp./fragments		1				1	2	6	
Ramiforms/fragments			2						

Section	San Giovanni 2											
Samples	06P-61	06P-62	06P-63	06P-64	06P-65A	06P-65B	06P-66	06P-67	06P-68	06P-69	06P-70	06P-71
dissolved amount	0.200	0.150	0.300	0.325	0.350	0.325	0.325	0.300	0.225	0.250	0.225	0.250
Icriodus alternatus alternatus										X	X	X
Palmatolepis crepida											X	
Palmatolepis falcata	X	X (cf.)			X							
Palmatolepis glabra glabra	X	X		X	X							X
Palmatolepis glabra pectinata		X	X			X	X	X				
Palmatolepis glabra prima			X	X	X	X	X	X	X	X	X	X
Palmatolepis gracilis gracilis			X (cf.)		X	X						
Palmatolepis minuta minuta			X	X	X	X	X	X	X	X	X	X
Palmatolepis minuta subgracilis												X
Palmatolepis minuta wolskae							X		X			X
Palmatolepis quadrantinodosalobata										X	X	X
Palmatolepis rhomboidea			X		X		X					
Palmatolepis schindewolfi					X	X	X					
Palmatolepis subperlobata							X (cf.)	X (cf.)	X	X	X	
Polygnathus glaber glaber					X	X	X					X
Polygnathus glaber bilobatus					X							
Polygnathus aff.glaber									X			
Icriodus sp.						X			X			
Palmatolepis sp./fragments	3		5	5	26	20	34	28	37	10	2	31
Polygnathus sp./fragments	1		1		5		1					2
Fragments/ramiforms	2	4	2	3	19	3	3	6	5	5	18	17

Section	Ponte Vina											Stilaro
Samples	06P-1	06P-2	06P-3	06P-4	06P-5	06P-6	06P-7	06P-9	06P-10	04A-1	04A-3	07P-257
dissolved amount	0.525	0.800	0.825	0.850	0.925	0.900	0.900	0.875	0.925	0.300	0.750	0.450
Ancyrodella buckeyensis	X											
Palmatolepis bogartensis		X										
Palmatolepis falcata			X									
Palmatolepis glabra distorta					X	X	X					
Palmatolepis glabra pectinata			X				X					X
Palmatolepis glabra prima			X			X						X
Palmatolepis gracilis gracilis			X		X		X				X	
Palmatolepis gracilis manca											X	
Palmatolepis gracilis sigmoidalis											X	
Palmatolepis marginifera marginifera			X	X	X							
Palmatolepis marginifera utahensis					X	X						
Palmatolepis minuta minuta						X						X
Palmatolepis minuta schleizia							X			X		
Palmatolepis schindewolfi			X					X	X	X	X	
Palmatolepis cf. uyenoi	X											
Polygnathus glaber bilobatus					X							
Polygnathus glaber medius			X	X								
Polylophodonta cf. confluens				X								
Scaphignathus velifer velifer									X			
Icriodus sp.				X								
Palmatolepis sp./fragments	16	6	>120	>120	38	42	22	6	12	13	10	3
Polygnathus sp./fragments	9	1	20	>80	13		7		3		30	
Fragments/ramiforms	42	12	72	>100	51	>100	11	52	>50	46	>100	7

Table III.1. Conodont elements abundance and distribution in the studied sections in the Stilo-Bivongi area.

III.2.1. General features and previous studies on the Paleozoic of the Stilo Unit in the area

In southernmost Calabria, the Stilo Unit crops out to the S and SE of the Aspromonte Massif, according to the general structure of the Southern Sector of the CPT (Fig. III.2.1). In this area, the Stilo Unit is mainly made up of Paleozoic metamorphites of the basament including all the lithologies present in the Serre Massif, but lacks Variscan magmatites (Serre and Cittanova granites) and high-grade metamorphic rocks.

The diverse outcrops of this unit in the Aspromonte can be grouped in two NNE-SSW discontinuous bands: Case Allai-Roccaforte del Greco-San Lorenzo (to the W) and Staiti-Pietrapennata-Palizzi Marina (to the E).

The more extensive one appears in the area between Case Allai, Roccaforte del Greco and San Lorenzo and corresponds to a cross-section through

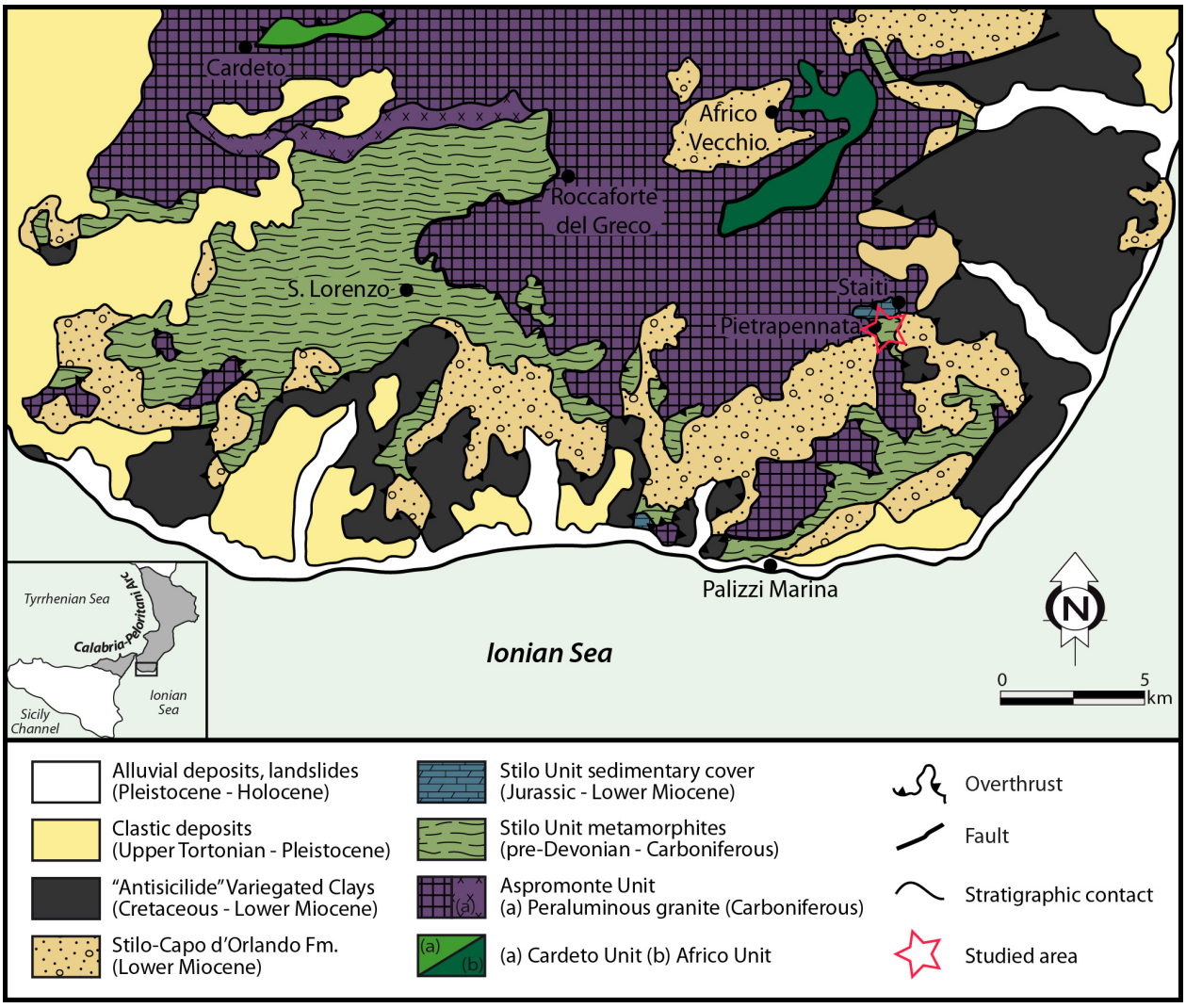


Figure III.2.1. Geological sketch of the southernmost Aspromonte (modified from Bonardi et al., 1982, after Graessner and Schenk, 1999).

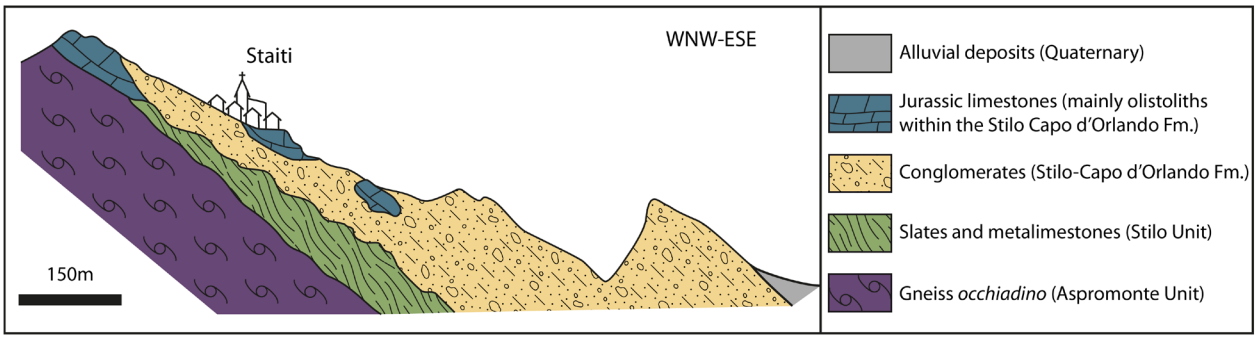


Figure III.2.2. Synthetic geological cross section in the Staiti region.

the upper continental crust, from the chlorite zone up to the sillimanite-muscovite zone (Graessner and Schenk, 1999). To the south, it is clearly covered by Stilo Capo D'Orlando Formation and the Argille Varicolori flysch deposits. The northern boundary with the Aspromonte Unit, however, is subject of debate. Crisci *et al*. (1982) and Bonardi *et al*. (1984b) delimited this contact as a normal fault between a broad band of basement gneisses of the underlying Aspromonte Unit and the schists of the Stilo Unit. Later, Graessner and Schenk (1999) pointed out the continuity of the metamorphic zonation on both sides of the boundary previously proposed and included an additional sillimanite-in isograd within the above mentioned metamorphic succession. Therefore, the Stilo Unit continues to the north and borders with peraluminous granites of the Aspromonte Unit.

The second group is formed by several little outcrops that appear between metamorphic rocks of the Aspromonte Unit and clastic deposits of the Stilo-Capo D'Orlando Formation. In particular, to the south of Staiti, the Stilo Unit crops out between the *gneiss occhiadino* of the Aspromonte Unit and the conglomeratic deposits of the Stilo-Capo d'Orlando Fm. (Fig. III.2.2). There, Paleozoic deposits of the Stilo Unit are mainly very weathered slates with metalimestone intercalations, locally thick in some intervals where three sections have been studied, measured and sampled for conodonts as I will describe in the next sections.

Previous detailed studies about the stratigraphy of the Stilo Paleozoic basement in this area are scanty and are commonly included in regional or petrographic papers. Crisci *et al*. (1982) and Bonardi *et al*. (1984) summarized the diverse opinions and correlated the *Unità superiore dell'Aspromonte* (upper Aspromonte Unit; Bonardi *et al*., 1979) with the Stilo Unit. They also pointed out their resemblance to the Stilo Unit in the Serre Massif.

Fossiliferous beds in the Stilo Paleozoic of the Aspromonte have been found only in their eastern outcrops. Gelmini *et al*. (1978) found dacryoconarids, conodonts, crinoids, echinoderms and gasteropods in diverse limestone intercalations in the Paleozoic phyllites, which they assigned to the Pietrapennata Unit. Later, Lorenzoni *et al*. (1980) corrected this assignation and confirmed that this succession actually belongs to the Stilo Unit. Two genus of dacryoconarids that they found, *Styliolina* sp. and *Novakia* sp., which range from Pragian to Frasnian (Lower-Upper Devonian). Among the conodonts they found (*Belodella devonica*, *Drepanodus* sp., *Hindeodella* sp., *Neoprioniodus* sp., *Plectospathodus* sp., *Spathognathodus* sp., and *Trichonodella* sp.), only *Neoprioniodus bicurvatus* has biostratigraphic implications and ranges from uppermost Silurian to the end of the Lower Devonian.

Bouillin *et al*. (1987) in a regional study on the Paleozoic successions of the CPT mentioned the presence of '*Ozarkodina*' *steinhornensis* near Staiti village, which indicates an Emsian age (Lower Devonian). Although they did not provide details on the location of the sample, it probably corresponds to that studied by Gelmini *et al*. (1978).

III.2.2. Studied sections

The two sections studied in the Aspromonte have been made in the Staiti-Pietrapennata area (FIg. III.2.X). As mentioned before, the rest of outcrops of the Stilo Unit in Aspromonte lack limestones intervals and show a high metamorphic grade. All three correspond to very thick limestone intercalations between metamorphites (slates and phyllites) of the Stilo Paleozoic.

From Staiti to Pietrapennata, along the *strada comunale* Ermanata-Staiti, in the left bank of the Torrente Aranghia, but next to its bottom, a mainly carbonate succession appears (Fig. III.2.3). The succession is made up of very thinly bedded limestones with slate partings (Fig. III.2.4.a and b) that show kink-bands and chevron folds that form two NE-vergent conjugate systems (Fig. III.2.4.c). The sequence strikes approximately NW-SE. Dacryoconarids are visible and very abundant. Several samples here taken have released Lower Devonian conodonts (see chapter III.2.3 for biostratigraphic details).

Along the same road, on the right bank of the Torrente Aranghia, the Paleozoic succession continues and is mainly made up of bluish slates. This metapelitic sequence includes a decimetre interval of strongly folded lydites, approximately 1 km to the NE of Pietrapennata. After few tens of metres of bluish slates, a thick limestone interval occur (Fig. III.2.3.C and III.2.4.d). This carbonate succession begins with few metres of bluish thin-bedded limestones, slightly deformed, with dacryoconarids, crinoids and orthoceratids (Fig. III.2.4.e). After a fault, very visible because of a colour change, the succession continues with very thin bedded marly limestones. In this last interval, two samples have yielded also Lower Devonian conodonts (see chapter III.2.3 for biostratigraphic details).

Next to Pietrapennata, along the same road but after the crossroad to Ermanata, another succession of yellowish metamarls between slates with platy limestone intercalation appears. Unfortunately none of the limestone samples taken in these intercalations have released conodonts.

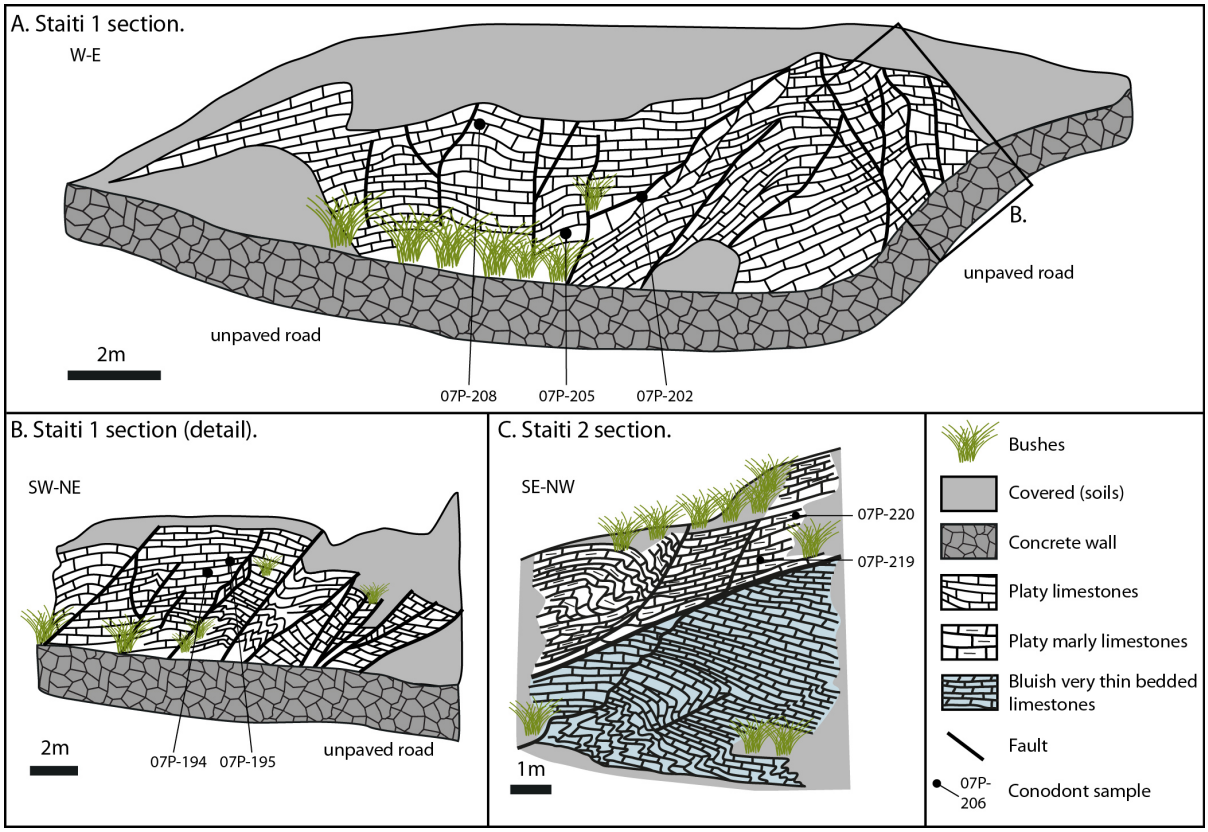


Figure III.2.3. Staiti 1 and 2 sections.

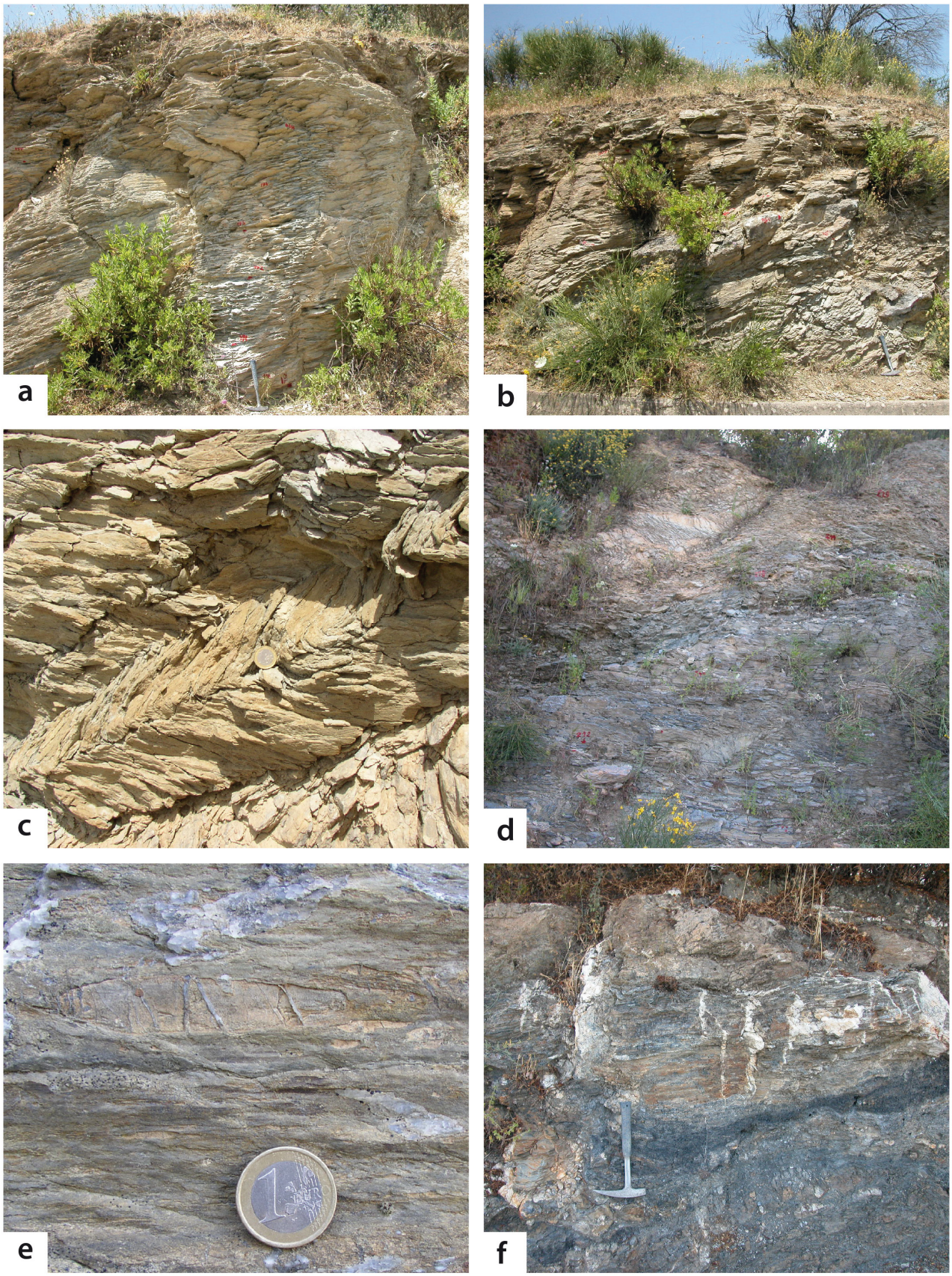


Figure III.2.4. Staiti 1 (a-c) and 2 (d-f) sections. ***a,*** *Lower interval of the Staiti 1 section with sample locations in red;* ***b,*** *Upper part of the Staiti 1 section;* ***c,*** *Detail of chevron folds in the lowermost part of the section;* ***d,*** *General view of the Staiti 2 section with samples locations in red;* ***e,*** *Orthoceratid section;* ***f,*** *Lydites, close to the Staiti 2 section.*

III.2.3. Conodont biostratigraphy

Two sections have been studied and sampled in the Paleozoic of the Stilo Unit in the Aspromonte Massif. A total of forty four samples were studied but only seven, which belong to the Staiti 1 and 2 sections, have released conodonts.

Five samples studied in the Staiti 1 section (fig. III.2.3.A and B) have released conodont elements, although they are broken and very but preserved. Several fragments and bad preserved Pa and Pb elements of *Icriodus* sp. Branson and Mehl, 1934 have been found in samples 07P-202, 07P-205 and 07P-208 (Fig. III.2.5.2 and 3), but they could not be classify at a species level. Sample 07P-192 released one Pa element of *Pseudooneotodus beckmanni* Bischoff and Sannemann, 1958 (Fig. III.2.5.1). As mentioned before, this species had a long-lasting biostratigraphic distribution (Middle Ordovician-Lower Devonian). Further studies and sampling are needed in order to be specific about the age of this succession.

In the Staiti 2 section (Fig. III.2.3.C), samples 07P-219 and 07P-220 have yielded several Pa elements of *Pseudooneotodus beckmanni* Bischoff and Sannemann, 1958 (Fig. III.2.5.4). As in the Staiti 1 section, further studies are needed.

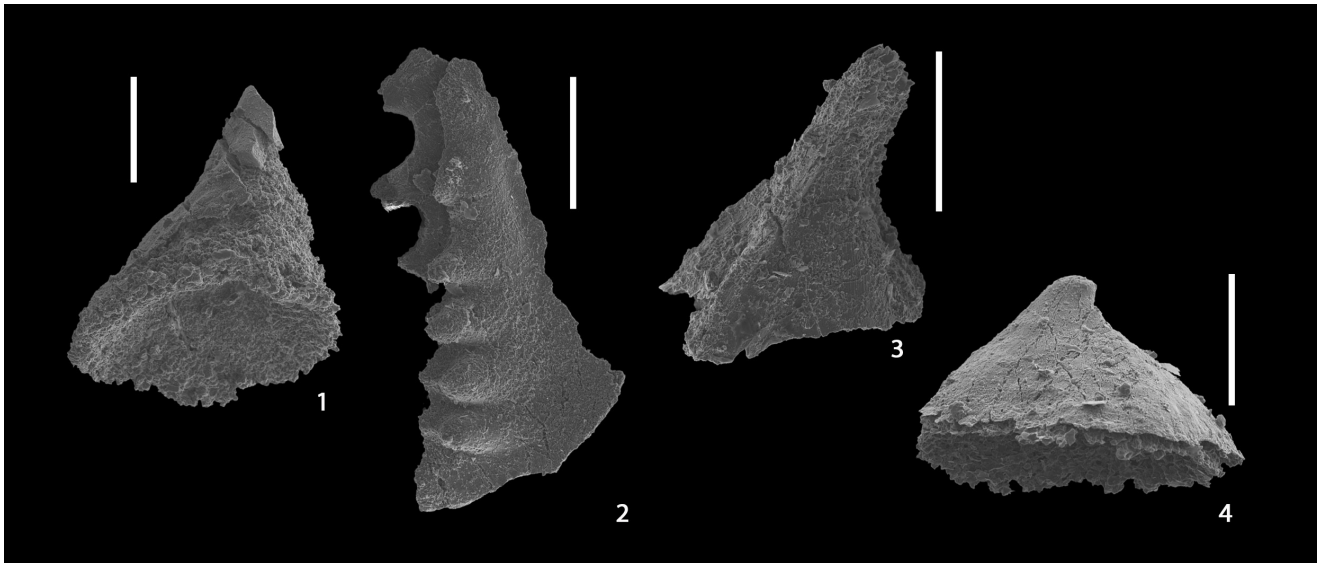


Figure III.2.5. Conodonts of the Staiti 1 (photos 1-3) and Staiti 2 (photo 4) sections. Scale bars are 200 µm.
***1,** Pseudooneotodus beckmanni; 07P-195-1.* ***2,** Icriodus* sp. *Pa element (fragment); 07P-202-1.* ***3,** Icriodus* sp. *Pb element; 07P-202-2.* ***4,** Pseudooneotodus beckmanni; 07P-220-1.*

III.3. Synthesis of the Stilo Unit Paleozoic stratigraphy

The correlation panel and conodont distribution chart shown in Figs. II.3.1 and II.3.2. summarize the litho- and biostratigraphy of the studied Paleozoic successions of the Stilo Unit. The main lithostratigraphic units recognized include: i) undated pre-Devonian (Cambro-Ordovician?) greenish-grayish metapelites with lenses of amphibolitic rocks and increasing metamorphic overprint downwards (Pardalà schists), which are topped by ii) low grade silvery-gray metapelites and metapsammites with isolated (metre-thick at most) limestone and black shale intercalations (Silurian?, Pardalà section); iii) Early to Middle Devonian dark gray pelagic limestones with pelitic intercalations, in some intervals nodular and containing orthoceratids and isolated corals, in addition to conodonts (Assi section); iv) gray pelites including thin limestone beds, locally containing dacryoconarids (San Giovanni 1 section) and dacryoconarids thin to medium-bedded limestones with some pelitic intervals (Staiti 1 and Staiti 2 sections; v) Upper Devonian, conodont-rich gray limestones, sometimes nodular and red, including isolated pelitic intervals downwards (S. Giovanni 1 and 2, Stilaro and Ponte Vina sections), within which the Upper Kellwassser event has been recognized; vi) Lower Carboniferous lydites (Ponte Vina section); and vii) Carboniferous Culm-like brownish pelites and sandstones.

The sedimentary protolith of the Pardalà schists was a thick siliciclastic (pelitic and fine-grained psammitic) succession including abundant bodies of basic volcanic rocks and vulcanoclastic deposits. This succession was probably deposited in not very deep marine environments under an extensional geodynamic regime that would correspond to the early stages of the evolution of a divergent continental margin. Persistent subsidence and basin isolations favoured deposition of black shales with Fe-Mn mineralized levels (ampelites) as well as of thin limestone intercalations in open marine episodically restricted environments with badly circulated water masses. Although this lithologic interval remains undated, a Silurian age can be reasonably inferred according to its stratigraphic position. Moreover, deposition of ampelites facies was related to some of the anoxic events accounted worldwide during the Silurian.

A pelagic carbonatic deposition, associated with fine-grained siliciclastic intervals characterizes the Devonian of the Stilo Unit. As mentioned above, significant vertical and lateral facies changes have been identified during this period (Fig. III.3.1). These must be interpreted in the frame of a changing hemipelagic to pelagic sedimentary basin within a faulted continental margin, with deposition of nodular limestones in outher shelf and high-bottom areas and of alternating pelagic limestones and pelites with local black shales in deeper areas. In addition, a deepening-upwards evolution can also be inferred from the Early Devonian to the Early Carboniferous.

As previously discussed by Navas-Parejo et al. (2009a), the Lochkovian–Lower Pragian to Eifelian orthoceratid limestones of the Fiumara Assi were deposited under pelagic conditions, with rare presence of shallow-water organisms. These conditions are typical of outer-shelf environments of moderate depth and of pelagic plateaus with condensed sedimentation (Holland et al., 1994; Histon and Schönlaub, 1999). The Lower-Middle Devonian pelites alternating with thin dacryoconarid limestones of the S. Giovanni 1 and of the Staiti 1 and 2 sections were deposited in moderately deeper environments than those where the Assi nodular limestones were deposited, as these successions contain much more pelitic intervals while limestone beds are less frequent (figs. III.3.1 and III.3.2).

During the Upper Devonian, equivalent depositional conditions as well as lateral facies can be inferred. In the S. Giovanni 2 section, Famennian deposits correspond with reddish nodular limestones that were probably deposited in a highbottom (pelagic swell) area. In the Ponte Vina section, coeval sediments were deposited

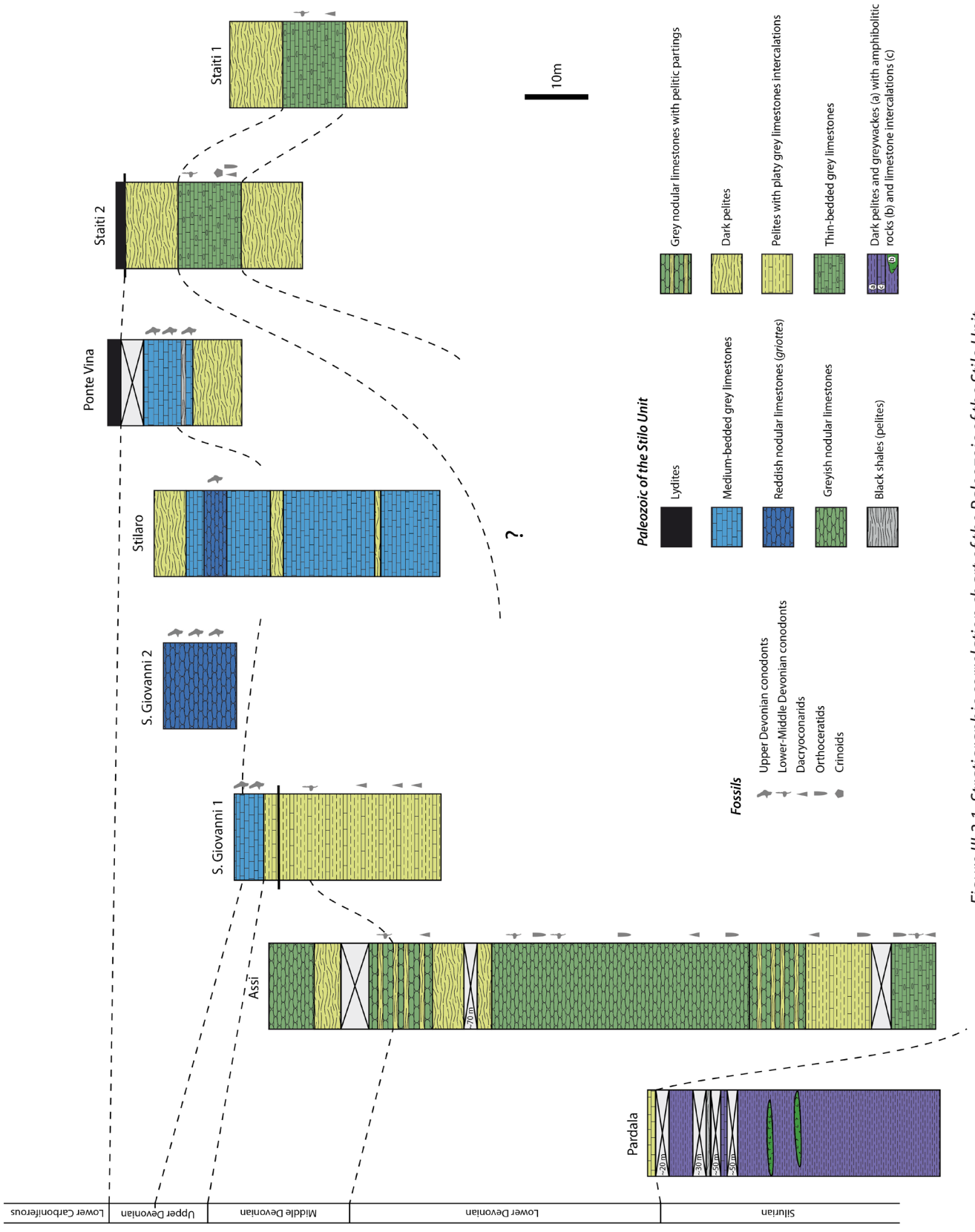


Figure III.3.1. Stratigraphic correlation chart of the Paleozoic of the Stilo Unit.

in slightly deeper environments, as they are characterized by greyish, pelagic, thin to medium bedded limestones. The Famennian limestones of the Stilaro section also include nodular facies slightly different to those of S. Giovanni and thicker than coeval beds of the Ponte Vina section, in spite of the short distance between them. This must be interpreted as due to the deposition of the Stillaro section in a transitional area between the San Giovanni 2 pelagic swell and the slightly deeper Ponte Vina depositional environment,

The outer position within the continental margin and the slightly deeper depositional setting inferred for the Ponte Vina section is reinforced by the presence there of the Upper Kellwasser anoxic event. This event is recorded by the thin black clay horizon observed there between Frasnian and Famennian beds dated by conodonts. Anoxic conditions developed in the Ponte Vina area at the end of the late Frasnian and disappeared during the Famennian. As previously discussed by Navas-Parejo et al. (2009b), this may explain why the Frasnian conodonts of the section studied are pyritized below the anoxic horizon whereas the Famennian conodonts in the bed immediately above it are not. Pyritization of the Frasnian conodonts was an early diagenetic phenomenon produced during burial under sulfidic conditions. As it is common throughout the world during the development of the Upper Kellwasser Event, this accounted during the deposition of the anoxic black pelitic horizon immediately above the dated Frasnian beds. After the anoxic event, the more or less complete reoxygenation of bottom waters made sulfidic conditions more difficult to be reached below the sediment-water interface and, consequently, the conodonts deposited within the limestones immediately above the anoxic event bed were rarely pyritized during early burial diagenesis. In addition, the palaeoceanographic changes related to the development of widespread anoxic conditions might also explain why the lower Famennian triangularis, crepida, and rhomboidea zones are not recorded in the Ponte Vina section. This stratigraphic gap of the lower part of the Famennian must be confirmed after more detailed studies, as the scarce conodont fragments extracted from the few-decimetre-thick limestone beds immediately overlying the black pelitic interval have not yet been classified.

In the Paleozoic record of the Western Mediterranean Alpine orogen, the Kellwasser crisis has until now only been recognized in the Carnic Alps and in the Malaguide Complex of the Betic Cordillera. In this latter region, Rodríguez-Cañero (1993b) obtained an Upper Devonian conodont fauna from limestone lenses, including several Fe-rich horizons testifying to Kelwasser Events and intercalated between Devonian-Carboniferous pelites and sandstones. Although no Kellwasser anoxic horizon has yet been detected in the Malaguide Complex, sea-level oscillations have been detected by the analysis of conodont biofacies (Rodríguez-Cañero, 1993b). The same has also been described in other areas (Schindler 1993; Crick et al. 2002; Bond et al. 2004).

Lithologically similar facies spectra and stratigraphic successions are frequent in the Silurian–Devonian record of other Mediterranean areas. In particular, the Stilo succession resembles those of SW Sardinia (Mason Porcus Fm.; Gnoli et al., 1988; Olivieri and Serpagli, 1990; Ferreti et al., 1998), eastern Pyrenees (Sanz-López, 2004), and other sites of the southern Variscan Chain (García-López et al., 1996). Equivalent coeval facies are also present in the Maláguide-Ghomaride Units of the Gibraltar Arc (Rodríguez-Cañero et al., 1990; Martín-Algarra et al., 2004). This is testifying to a common palaeogeographic origin for most of these domains.

In the Alps (Schönlaub, 1996; Schönlaub and Histon, 1999) equivalent successions are interpreted as related to the rifting and later drifting of North-Gondwanic fragments, leading finally to the opening of the Palaeotethys Ocean during Silurian–Devonian (von Raumer et al., 2003). A derivation from the northern Paleotethyan margin has been proposed for most Western Mediterranean Paleozoic successions (Stampfli and Borel, 2004). The Stilo succession and equivalent Paleozoic continental margin successions of the Calabria-Peloritani Terrane and of their counterparts in the Western Mediterranean Alpine belts also made part of the western Paleotethys. Nevertheless, the much more common presence of basic volcanics in the pre-Devonian succession of the Stilo Unit points to a palaeogeographic position closer to the rift axis related to the opening of the Paleotethys than that of the Malaguide successions, where basic volcanic rocks are much less common (apart from a system of dykes that are usually interpreted to be of Alpine age).

The palaeogeographic location proposed for the Stilo Unit and for the Malaguide and equivalent successions within a Paleothethyan margin is very different to that commonly accepted for the Variscan

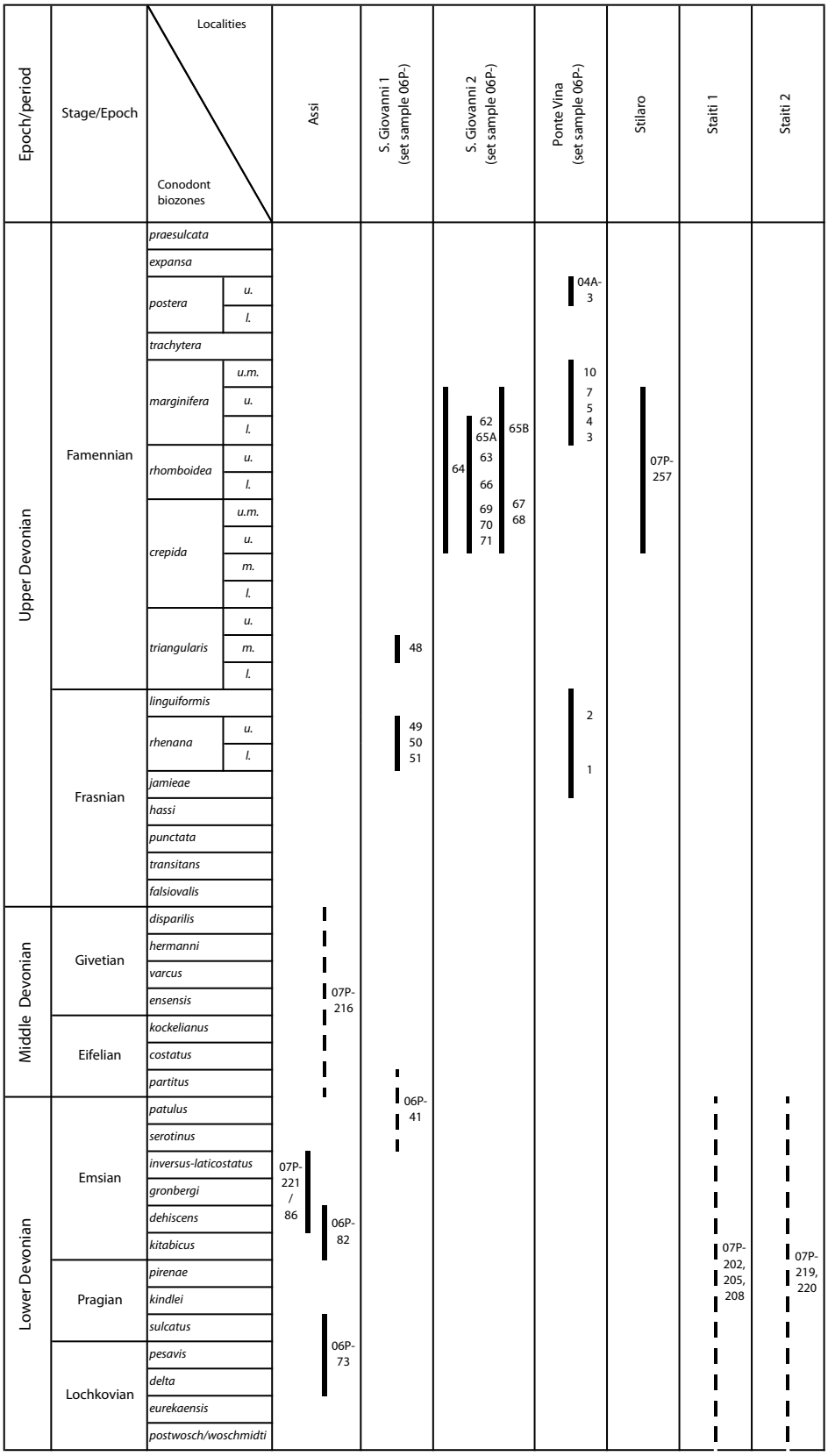


Figure III.3.2. Biostratigraphic results in the Paleozoic of the Stilo Unit.

regions of the Iberian Massif and of the Moroccan Meseta. These European and North-African Variscan regions come from areas palaeogeographically located far towards the west of that of the Alpine-Mediterranean regions. They are usually related to the Rheic Ocean and, in particular to its North-Gondwanic paleomargin (Simancas et al., 2005). An important difference between these Variscan regions and the intra-Alpine Paleozoic terrains (in the sense of Stampfli, 1996) is related to the different Devonian-Carboniferous histories of both areas: the Rheic Ocean was completely subducted during the Devonian–Early Carboniferous, whereas subduction of the westernmost Paleotethys was accomplished later, in Late Carboniferous-Permian time (Stampfli and Borel, 2004).

During Early Carboniferous time the Paleotethyan paleomargin where the Paleozoic successions of the Stilo Unit were deposited underwent a maximum deepening. This is testified by the presence of lydites above well-dated Famennian limestones. It certainly accounted during the mature passive margin stage dominated by thermal subsidence, which made the margin much deeper and topographically homogeneous. From this time on, the up to now divergent paleomargin became an active margin as it is indicated by the deposition of pelites and rare greywackes with Culm-like facies.

Mid-Carboniferous siliciclastic sediments of the Stilo Unit are traditionally considered synorogenic (Variscan) Culm (-like) deposits, as in the Malaguide Complex and other peri-Mediterranean areas. However, in the Stilo–Bivongi area, where this succession is best documented, the Culm-like facies are much more fine-grained and include rare conglomeratic beds unlike the coeval beds of the Malaguide Complex, where coarse-grained and thick conglomerate bodies are either very characteristic (Almogía Fm.) or predominant (Marbella Fm: see previous chapter). This suggests a more distal location of the Carboniferous Stilo-Bivongi succession when compared to coeval sections of the Malaguide Complex (especially, that of the Upper Malaguide Unit), or to its deposition within a tectonically less unstable part of the basin.

III.4.1. Previous studies on the Paleozoic stratigraphy and volcanism of the Longi-Taormina Unit

The Longi-Taormina Unit (hereafter LTU) crops out along a WNW-ESE trending belt, stretched from Sant'Agata di Militello in the Tyrrhenian coast to Taormina in the Ionian coast. Since the LTU successions, mainly in the central area, are covered by late-orogenic, wedge-top basin deposits (Stilo-Capo d'Orlando Fm.), their outcrops are discontinuous and are divided in three sectors: western (from Acquedolci to Banco di Formisia-Pagliara: figs. III.4.1 and III.4.2), central (from Pizzo del Moro to Malvagna) and eastern (from Novara di Sicilia to Taormina; Fig. III.4.2).

As mentioned before, the LTU consists of several subunits. These are independent tectonic slices (Fig. III.4.4.a) that are distinguished according to their geographic and tectonic location, and to the different sedimentary evolution of their Meso-Cenozoic covers. These LTU subunits have been differently named in the literature (Duée, 1961; Truillet, 1962, 1969/70; Caire *et al*., 1965; Lentini, 1975; Lentini and Vezzani, 1975; Bonardi *et al.*, 1976; Arnone *et al.*, 1979 ; Bouillin *et al*., 1992). Following Somma (2006) the entire LTU can be divided in three main subunits, denoted, from base to top as Lower (Capo Sant'Andrea-Gallodoro; figs. III.4.1 and III.4.2), Middle (Taormina; figs. III.4.1 and 2) and Upper (San Marco d'Alunzio-Rocca Licopeti; Fig. III.4.1) Subunit.

The Paleozoic successions in different imbricated subunits of the LTU are quite similar. They mainly consist of an alternation of metasiliciclastic rocks (metapelites, metasiltites, and metarenites with quartzite or metaconglomerate intercalations), locally with metacarbonate beds, and both basic and acidic metavolcanites (figs. III.4.4 and III.4.6).

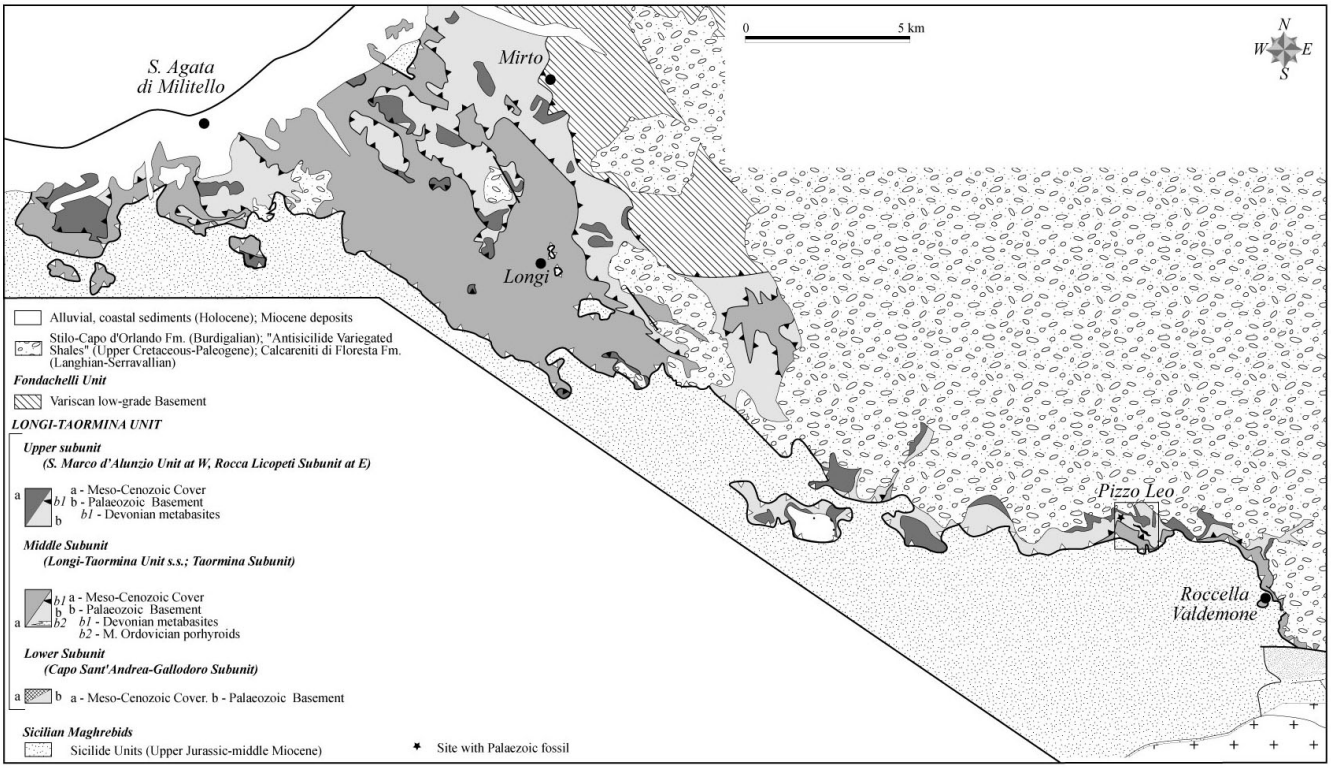


Figure III.4.1. Tectonic map of the Longi-Taormina Unit in the western and central outcrops.

In spite of ductile deformation and metamorphism due to its lower tectonic position in the nappe stack, the pre-Triassic rocks of the LTU still preserve their original Paleozoic stratigraphy. However, the succession is reconstructed from scanty, scattered and poorly constrained biostratigraphic data, as few fossiliferous layers have been found up to now: acritarchs testify for Cambrian-Ordovician ages; tentaculites and conodonts testify for to Devonian ages (Lardeux and Truillet, 1971; Majesté-Mejoulas *et al.*, 1986; Bouillin *et al.*, 1987; Acquafredda *et al.*, 1991). Previous stratigraphic syntheses were proposed by these authors using outcrops of different localities of the central-eastern Peloritani Mts. Majesté-Mejoulas *et al.* (1986) reconstructed a practically continuous from Cambrian-Ordovician to mid-Carboniferous succession, starting with pre-Upper Cambrian (?) phyllites, and followed upwards by Cambrian-Ordovician metapelites and metasandstones (Fig. III.4.4.b and c), probably Silurian (?) ampelites and thin metalimestones, Devonian shales, metasandstones, metalimestones and calc-schists, and Carboniferous (?) black radiolarian metacherts, metarenites and microconglomerates. The succession reconstructed by Spalletta and Vai (1989) recognized uppermost Cambrian-Arenigian (?) slates and sandstones with calc-alkaline volcanics below a Lower Devonian-mid Carboniferous (?) succession consisting of alkaline basic volcanics, pelagic limestones, slates, cherts, sandstones and conglomerates. Later, Acquafredda *et al.* (1994) proposed a synthetic stratigraphic column, spanning from Cambrian-Ordovician to Carboniferous and substantially similar to that of Majesté-Menjoulas *et al.* (1986), but characterized by a great amount of acidic (below) and basic (above) volcanic rocks, ranging from Early Ordovician to Early Carboniferous.

Paleozoic volcanism in the LTU has been documented since Truillet (1968), Duée (1969) and Atzori (1970). Many studies point out the presence

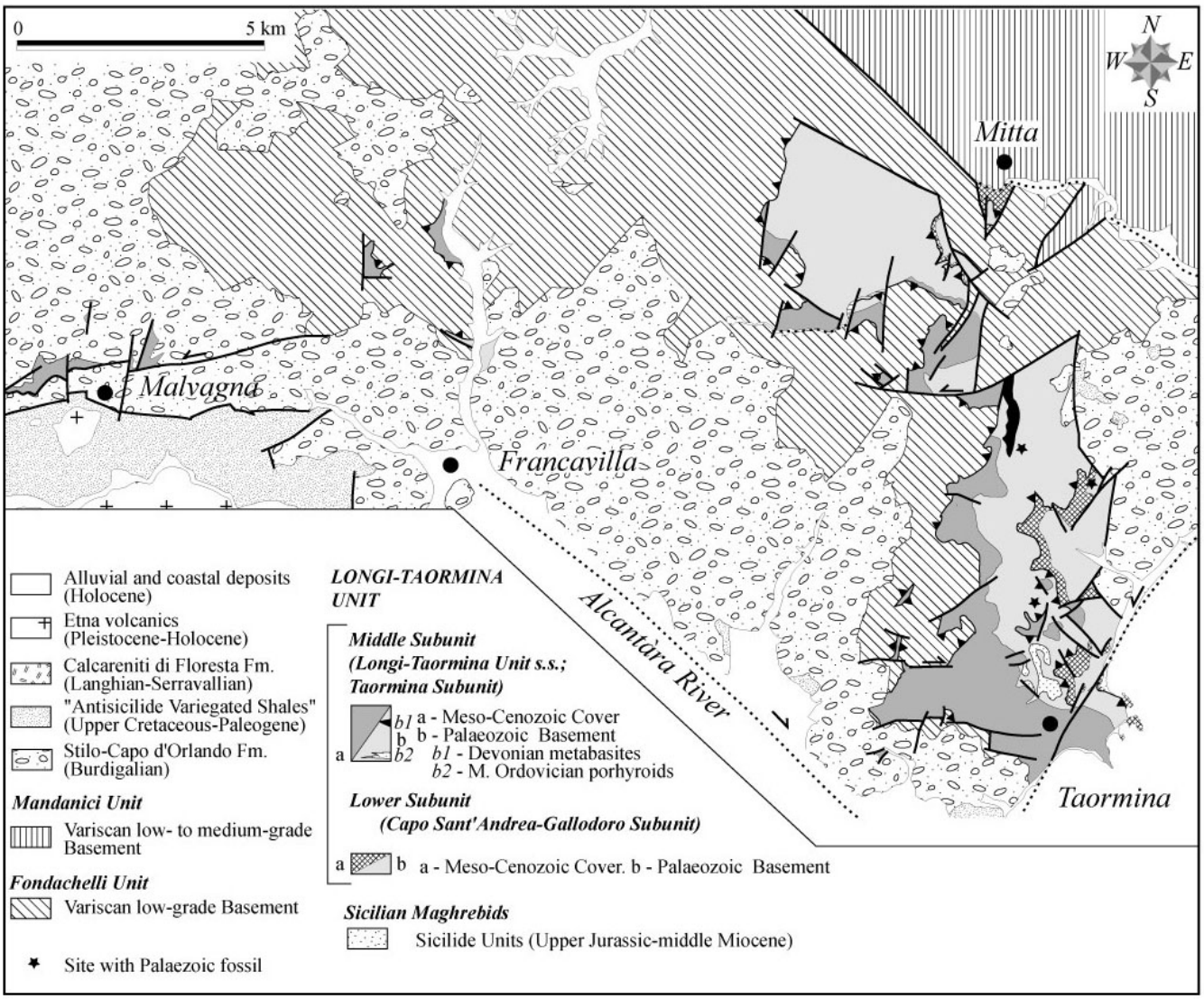


Figure III.4.2. Tectonic map of the Longi-Taormina Unit in the eastern outcrops.

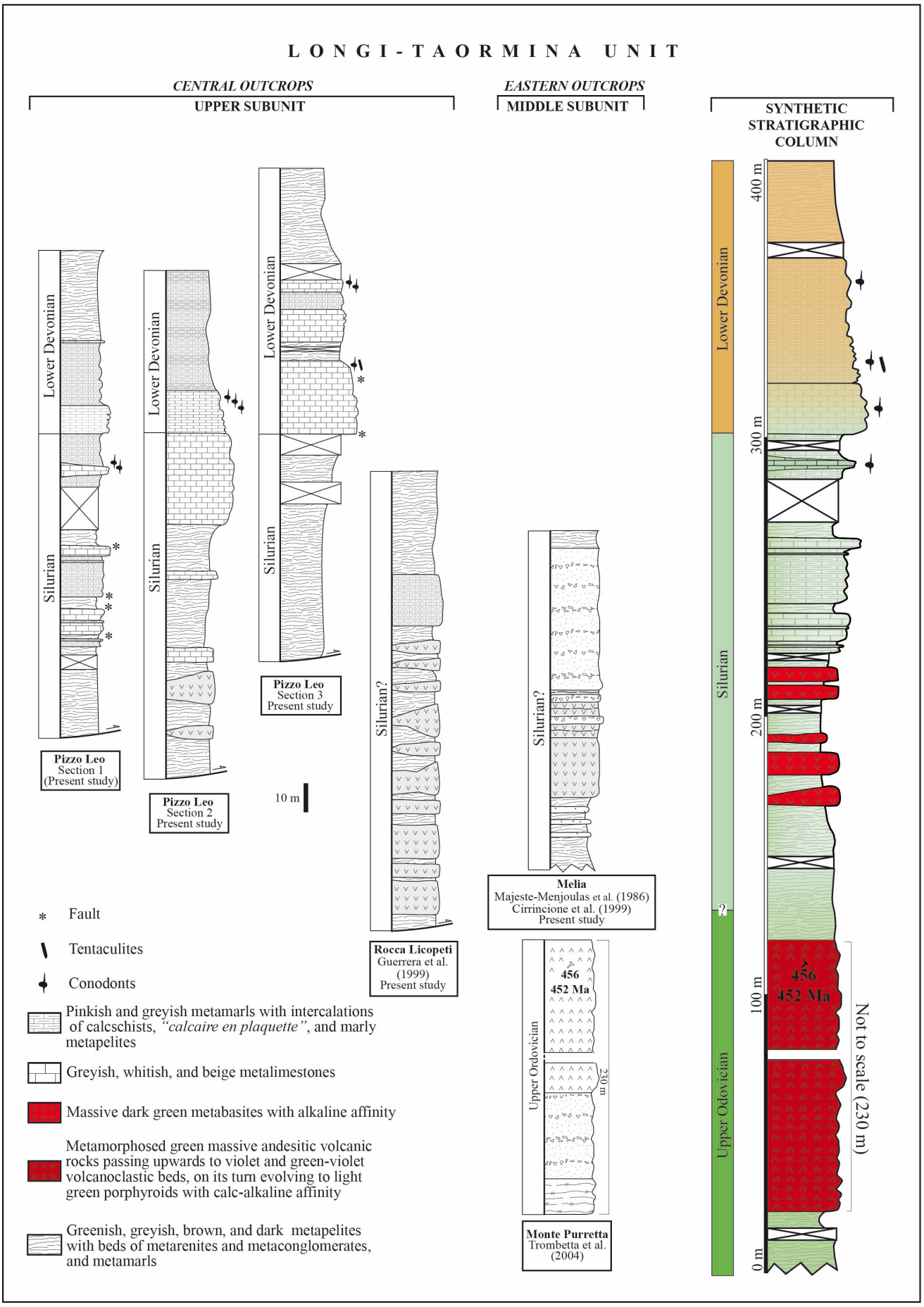


Figure III.4.3. Stratigraphic correlation panel of the Longi-Taormina Unit.

Figure III.4.4. ***a,*** *General overview of the LTU lower and middle imbricates (subunits) in the Taormina-Castelmola area, with the Paleozoic basement (Pz) unconformably covered (dashed line) by Triassic Pseudoverrucano (PsV) and Jurassic carbonates (J);* ***b,*** *Outcrop of Ordovician acidic volcanites along the Taormina Castelmola road, Middle Subunit;* ***c,*** *Close up of the acidic volcanites in the same site as (b); the picture is around 20 cm wide;* ***d,*** *Strongly folded succession of metabasites outcropping along the road from Letojanni to Roccafiorita; e, Close-up of the area squared in (d) showing a microduplex structure formed by a thin quartzite horizon between two thicker beds.*

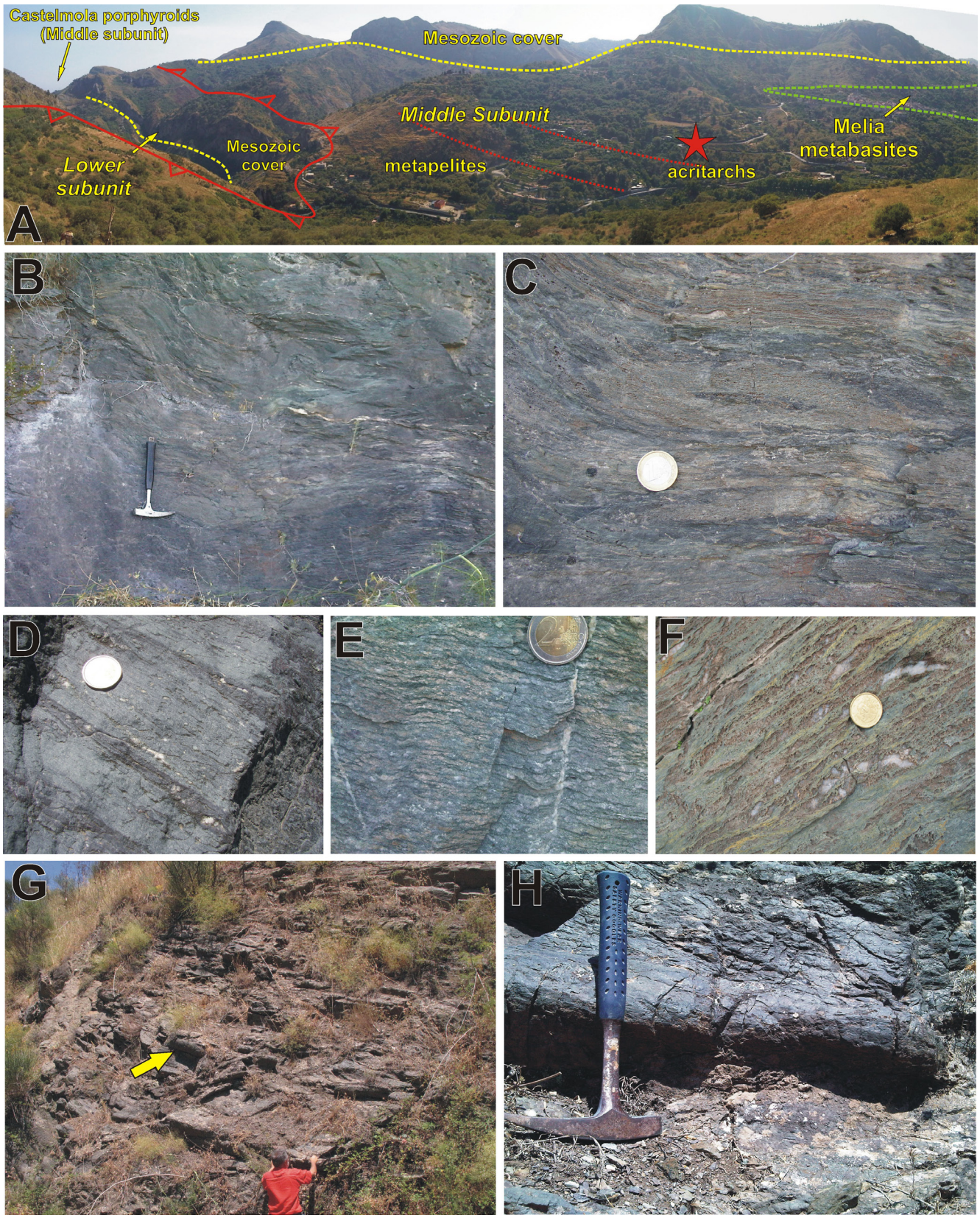


Figure III.4.5. Metabasites of the Middle Subunit (a to f: Mongiuffi-Melia area) and Upper Subunit (g to h: Rocella Valdemone area) of the LTU. ***a,*** *Panorama of the lower and middle subunits of the LTU in the sector from Castelmola to Melia (N towards the right). The dashed yellow line line indicates the unconformity at the base of the Mesozoic cover; the dotted line indicates the regional dip of the foliation of the Paleozoic metapelites of the Middle Subunit;* ***b,*** *Melia metabasites, with flattened and sheared pillows;* ***c,*** *Close up of sheared pillow metabasalts of the same outcrop as (b);* ***d,*** *Vacuolar metabasalt (e) Strongly sheared metabasalt;* ***f,*** *Metahyaloclastite;* ***g,*** *Flattened pillow metabasalts of the Upper Subunit in the Rocca Licopetti area;* ***h,*** *Close-up of the pillow arrowed in (g).*

of three main groups of volcanites (Ferla, 1978; Ferla and Azzaro, 1978; Atzori and Ferla, 1979; Atzori *et al*., 1982, 1982-83, 1984; Ferla *et al*., 1982-83; Acquafredda *et al.*, 1991, 1992, 1994; Cirrincione *et al*., 1999, 2005; Guerrera *et al*., 1999; Trombetta *et al*., 2004; Messina and Macaione, 2010). The first group consists of calc-alkaline volcanics of orogenic affinity, composed of andesites, dacites, rhyodacites and rhyolites, with wide outcrops in the Middle Subunit of the eastern sector (Castelmola area: figs. III.4.2 and III.4.4.a and e-f). Recent U-Pb and geochemical data (Trombetta *et al*., 2004) have demonstrated the Ordovician age (456–452 Ma) and calc-alkaline affinity of the Castelmola volcanic suite (as previously suggested by Spalletta and Vai, 1989, and Acquafredda *et al*., 1992).

The second group of volcanites, occurring in all subunits, is characterized by submarine eruptions, responsible for alkaline (locally pillow) basalts and dolerites (Fig. III.4.6.b-e), as well as for volcanoclastic products (tuffs and hyaloclastites: Fig. III.4.6.f). Their widest and thickest outcrops are located in the western (Caprileone area) and central (Roccella Valdemone area) sectors of the Upper Subunit (Fig. III.4.1), and in the eastern sector of the Middle Subunit (Mongiuffi-Melia area: Fig. III.4.2). Geochemical data from these volcanites correspond to intraplate basalts and testify for intra-continental rifting (Ferla, 1978; Ferla and Azzaro, 1978; Atzori *et al*., 1984, 2001; Acquafredda *et al*., 1991, 1992; Cirrincione *et al*., 1999, 2005; Trombetta *et al*., 2004). Contrasting datings have been proposed on these alkaline metabasalts, however. Those outcropping in the Mongiuffi-Melia area of the Middle Subunit (Fig. III.4.5.a-f) were interpreted to be Devonian (Ferla, 1978, 2000; Ferla and Azzaro, 1978; Atzori and Ferla, 1979; Atzori *et al*., 1982-83; 1984; Ferla *et al*., 1982/83), whereas a Cambrian-Ordovician age was proposed by Bouillin *et al*. (1987) on the base of acritarchs found in green metapelites and metarenites below the volcanites (Fig. III.4.5.a). The latter age has been used in some recent palaeogeographic and geodynamic reconstructions (Atzori *et al*., 2001; Cirrincione *et al*., 2005). Nevertheless, the metabasites with alkaline affinity found in the Upper Subunit exposed in the Rocca Licopeti area, close to Roccella Valdemone (figs. III.4.1 and III.4.5.g-h), have been unanimously considered to be of Devonian age because they are directly overlain by calc-schists and limestones with tentaculites (Ferla, 1978; Atzori *et al*., 1982/83; Ferla *et al*., 1982/83; Atzori *et al*., 1984; Guerrera *et al*., 1999).

The third group of volcanites is represented by Carboniferous (?) arc-tholeiites to dacites recognized in the Upper Subunit of central and western sectors (Ferla, 1978; Ferla and Azzaro, 1978; Bouillin *et al*., 1987; Acquafredda *et al*., 1991; 1992; 1994). However, no isotopic or biostratigraphic data support the Carboniferous age suggested for these calc-alkaline vulcanites.

In conclusion, the only documented dating for metavolcanic rocks of LTU is the Ordovician age of the calc-alkaline to rhyolitic volcanites close to Castelmola. The age of the other magmatic events, on the contrary, is not well constrained.

III.4.2. Studied sections

The available data on the LTU Paleozoic according to previous stratigraphic synthesis and studies on volcanism have been critically revised, completed and updated with new field observations, and with sampling for conodont biostratigraphic studies. The results presented in sections III.4 andIII.5 have been recently submitted as a scientific paper prepared in collaboration with Dr. Somma and my supervisors, Dr. Martín-Algarra, Dr. Rodríguez-Cañero and Dr. Perrone.

Field observations and conodont sampling were accomplished in all the LTU three subunits. Six sections have been measured, sampled and studied for conodonts in the Taormina-Castelmola, Gallodoro and Mongiuffi-Melia areas (figs. III.4.1 and 2), and particularly in the Roccella Valldemone and Pizzo Leo areas (Fig. III.4.6) where biostratigraphic information has been obtained.

III.4.2.1. Lower Subunit (Capo Sant'Andrea-Gallodoro Subunit)

The Lower Subunit corresponds to the Longi-Gallodoro Unit of Caire (1970), the Capo Sant'Andrea Unit of Lentini and Vezzani (1975), the A and B *p.p.* Units of Bonardi *et al.* (1976). Its succession crops out only in the eastern sector (Fig. III.4.2), where it is mainly exposed in a narrow belt, stretched from Gallodoro to Capo Sant'Andrea along the Letojanni and Gallodoro streams, and south and SW of Mitta village. The Paleozoic succession reaches a maximum thickness of 150 m and it is mainly composed of undated metamorphosed siliciclastic rocks, including greenish-greyish metapelites and metapsammites with metaconglomerate, quartzite layers and small metacarbonate lenses. As concerns volcanic rocks, porphyroids, very likely reliable to similar Upper Ordovician rocks, widely cropping out in the Middle Subunit, and alkaline basalts, have been recognized in little outcrops (Ferla, 1978). In the upper part of the succession, 0,5 km to the south of Gallodoro, the metapelites and metapsammites are locally overlain by grey metalimestones and slightly metamorphosed nodular limestones alternating with green metapelites. In these metalimestones Majésté-Menjoulas *et al.* (1986) found conodont fragments that they attributed to *Polygnathus communis communis*, of Late Devonian age. The six samples studied from this outcrop, however, did not provide conodont remains, but only abundant crinoid ossicles.

III.4.2.2. Middle Subunit (Taormina Subunit)

The Middle Subunit is equivalent to the Longi Nappe of Ogniben (1960), the B *p.p.* and C Units of Bonardi *et al.* (1976), and the Longi-Taormina Unit *s.s.* of Lentini and Vezzani (1975). It is the widest outcropping subunit and it is spread from the western (Sant'Agata di Militello: Fig. III.4.1) to the eastern sector (Taormina: Fig. III.4.2). The Paleozoic rocks of the Middle Subunit are mainly exposed in the eastern sector, where they show a thickness of about 250-300 m (figs. III.4.4.a and III.4.5.a). Because of tectonic elision, the thickness and outcrop area of these rocks decrease westwards, up to almost completely disappear in the western coast, where this subunit is mainly represented by the Mesozoic-Cenozoic cover (Fig. III.4.1). In the eastern area, the outcrops of Fondachelli Unit, which thrusts directly over the Middle Subunit, are characterized by fine-grained black schists very rich in graphite; these black schists remember some rocks interbedded with the metabasites of the Middle Subunit in this area. Similar rocks near Antillo were interpreted by Majésté-Menjoulas *et al.* (1986) as Silurian meta-ampelites (former black shales) but our sampling of metalimestone lenses associated with these black metapelites in search for conodonts was unsuccessful.

The best outcrops of the Paleozoic succession of the Middle Subunit can be observed along the Taormina-Castelmola road (Somma *et al*., 2005b), and in the Mongiuffi-Melia and Galfa Mt. areas (figs. III.4.4.a and III.4.5.a). It is mainly made up of greenish-greyish metapelites and metapsammites with metre-thick beds of quartzites and interbedded with volcanic rocks. Whitish to light yellowish-greenish, intermediate to acidic metavolcanics, up to 100 m-thick, lie on the metapelites, and are locally associated with sub-alkaline metatuffs in the Monte Purretta-Castelmola area, close to Taormina: Fig. III.4.5b-c). As mentioned above, this Castelmola volcanic suite was dated as Late Ordovician by Trombetta *et al*. (2004). N of Castelmola, near Ogliastrello, these acidic to intermediate volcanic rocks are overlain by dark coloured metapelites. In the area close to Portella Pandolfo, in a limestone lens associated with these metapelites, Bouillin *et al.* (1987) found undeterminable but probably late Devonian conodont fragments below a few metres thick succession of well bedded dark cherts (lydites). We were unable to find these limestones and, consequently, to sample them, but we have found the lydites, which show a very similar lithofacies to that of well-bedded radiolarian cherts overlying Fammenian limestones dated by conodonts in the Stilo Unit (Navas-Parejo *et al.*, 2009b). To the N, the outcrops of the Middle Subunit Paleozoic are dominated by dark-coloured metapelites and metapsammites with unknown thickness.

Field observation and mapping suggests that the calc-alkaline volcanic rocks and associated metapelites outcropping in the Castelmola area constitute the lower part of the succession, although their geometric relationships with those

outcropping to the N in the same unit are not fully understood. Actually, the metapelite outcrops bearing calc-alkaline volcanites is cartographically continuous toward the N, up to the Mongiuffi-Melia (Fig. III.4.5.a) and Galfa Mt. areas, but the complicated tectonic structure renders difficult to recognize the stratigraphic polarity of the succession and strong tectonic disturbance can be observed in the metabasites included within this part of the succession (Fig. III.4.5d-e). Within the greenish-greyish metapelites and metapsammites outcropping to the S of Melia, Bouillin *et al.* (1987) found Cambrian-Ordovician acritarchs (Fig. III.4.5.a). These metapelites certainly underlie the Mongiuffi-Melia alkaline basalts (metabasites *s.s.*: Ferla and Azzaro, 1978), and are sometimes pillowed (Fig. III.4.5.b-c) and vacuolar (Fig. III.4.5.d) but usually strongly flattened and sheared (Fig. III.4.5.e). They also are locally associated with, and followed by, violet to greenish metatuffs (known as varicoloured schists: Fig. III.4.5.f). These metabasites can reach a thickness of 50 m.

III.4.2.3. Upper Subunit (San Marco d'Alunzio-Rocca Licopeti Subunit)

The Upper Subunit corresponds to the Galati Nappe of Ogniben (1960), the San Marco d'Alunzio Unit of Lentini and Vezzani (1975), the B *p.p.* Unit of Bonardi *et al*. (1976) and the Rocca Licopeti Unit of Bouillin *et al*. (1992). It is is exposed in the western and central sectors, from Sant'Agata di Militello to Rocca Licopeti (Fig. III.4.1), where it reaches a thickness of 300 m. The outcrop extent decreases towards ESE, and the subunit completely disappears in the eastern sector (Fig. III.4.2). The succession is fairly similar to that characterizing the Middle Subunit. It mainly consists of greenish-greyish metapelites and metapsammites, with minor metalimestone beds, and includes some layers of acidic and basic metavolcanics both in the western and central sectors (Duée, 1961; Ferla, 1978; Atzori *et al*., 1984).

Alkaline metabasites are widespread and predominant in the Paleozoic succession of the western sector, but good outcrops are rare. The best outcrop of these metabasites is located in the central sector, at Rocca Licopeti, about 2 km NNE of Roccella Valdemone (Fig. III.4.1). There, alkaline metabasalts locally preserving pillow structures (Fig. III.4.5.g-h) lie on volcanic metabreccias (about

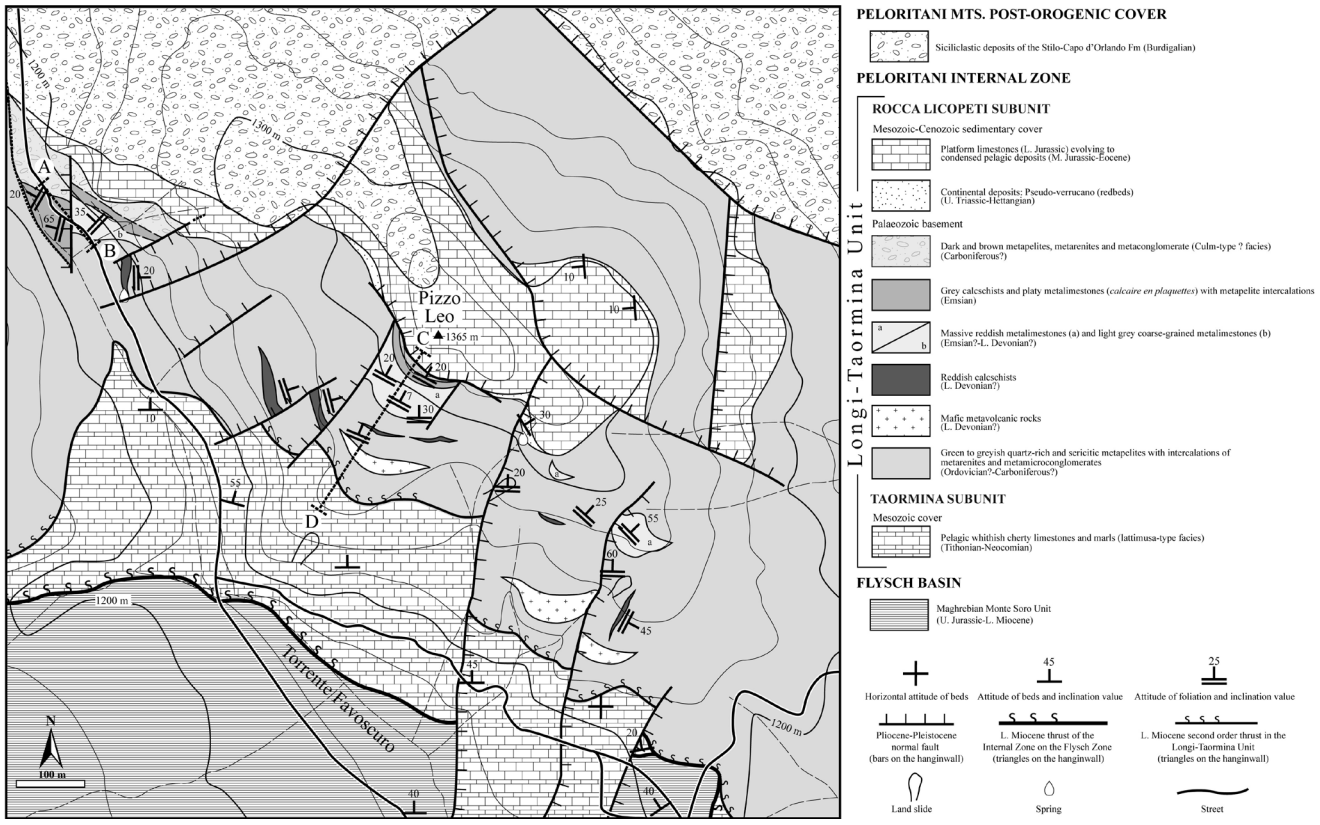


Figure III.4.6. Geological map of the Pizzo Leo area.

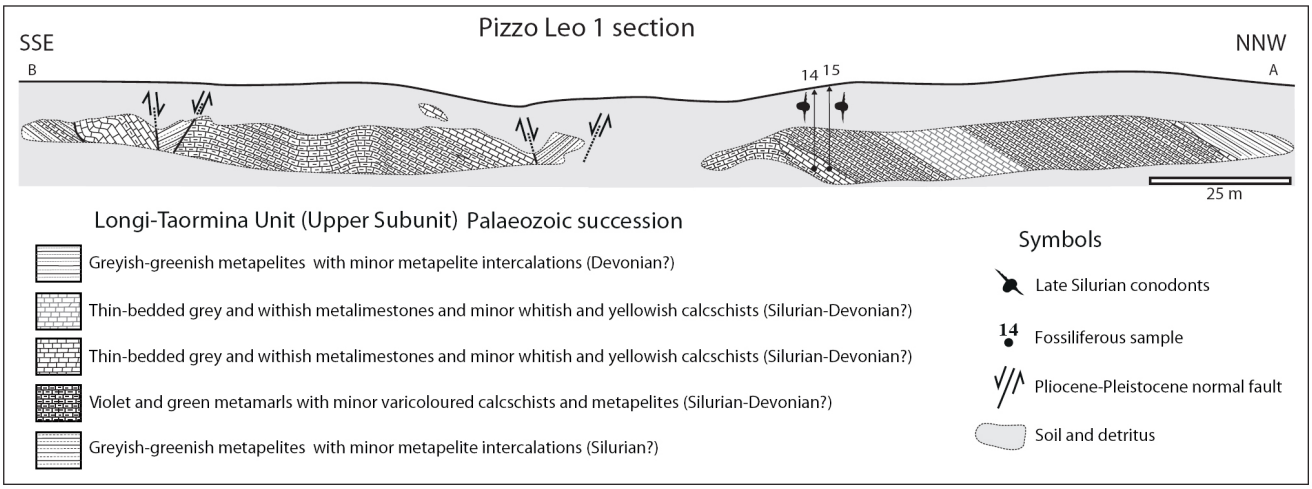


Figure III.4.7. Pizzo Leo 1 geological section (see location of the section in Fig. III.4.6).

1 m thick) that rest on metapelites. The metabasalts are overlain by metatuffs (0,5 m thick) that underlie a 6 m-thick level of calc-schists (Guerrera *et al*., 1999). Similarly, in the Portella Sant'Antonio area, 250 m west of Licopeti stream, the metabasites are overlain by a metacarbonate succession, up to 15 m-thick, composed of strongly folded light grey calc-schists followed by pinkish metamarls (Somma, 1998). In the metapelites associated with the metabasites of the western sector, Duée (1961) found thick-bedded metalimestones lenses, up to some tens of metres thick, which he interpreted as possible Silurian facies. However, biostratigraphic data in this sector are limited to dacryoconarids (*Styliolina*), spanning from Pragian to Famennian (Devonian), in one arenaceous metalimestone layer outcropping in the Fitalia stream (Acquafredda *et al.*, 1991). We were unable to find these metalimestone beds.

More biostratigraphic data are available in the central sector near Pizzo Leo, about 5 km SE of Floresta; Fig. III.4.6), where the Paleozoic succession of the Upper Subunit has been studied in detail in three outcrops (see below). The abundance of intercalated metacarbonate layers within the metamorphosed siliciclastic succession in this area was first revealed by Lardeaux and Truillet (1971), who found one limestone bed rich in tentaculites near to the base of a pelagic limestone succession cropping out along the Randazzo-Floresta national road 116 NR between km 12.7 and 12.8. Among dacryoconarids, they recognized *Styliolina* and presumed forms of *Viriatellina*. This association indicates a Pragian to Givetian age (Lower-Middle Devonian). Later, Majesté-Menjoulas *et al*. (1986)

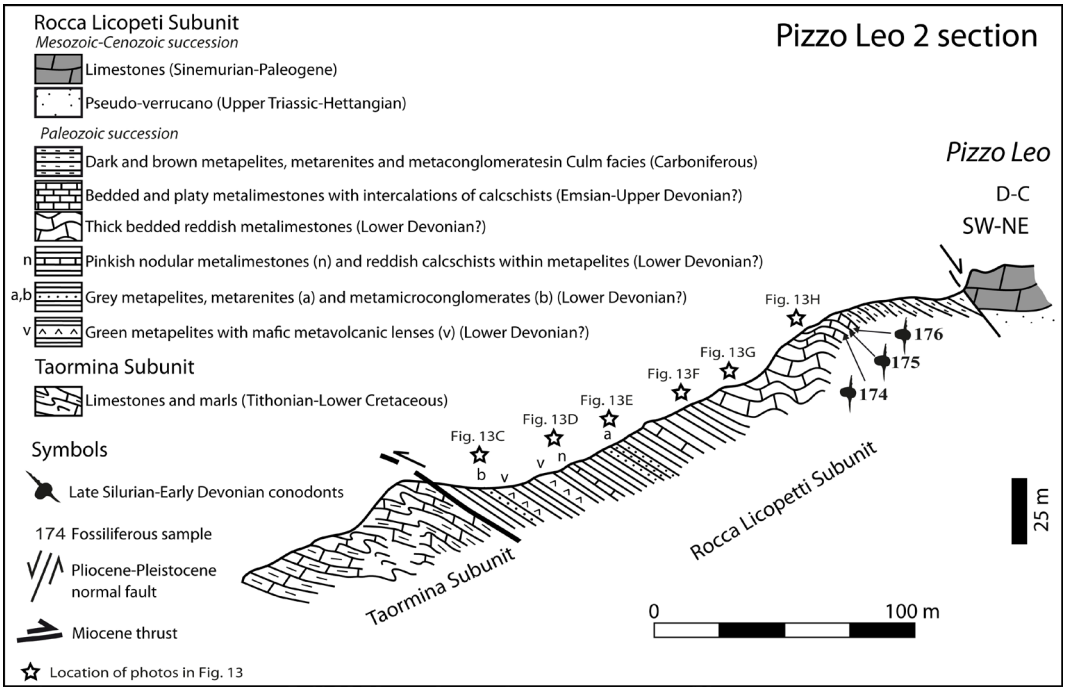


Figure III.4.8. Pizzo Leo 2 geological section (see location of the section in Fig. III.4.6).

studied a several tens of metres thick metalimestone succession along the Favoscuro stream, running parallel to the 116 NR. They recognized, from bottom to top:

a) violet metapelites with marly layers, gray to pink calc-schists and limestone intercalations bearing badly preserved Devonian conodonts (*Pseudopolygnathus* sp., *Bispathodus stabilis*, and *Bispathodus* sp.; and *Polygnathus* sp. in an upper horizon).

b) a few metres of black radiolarian cherts (lydites) that they attributed to the Carboniferous, on the base of facies comparisons with equivalent rocks outcropping in the Middle Subunit (Portella Pandolfo area: Fig. III.4.4.d) and, above Upper Devonian metalimestones dated by conodonts, in the Serre-Aspromonte massif (Stilo Unit).

c) a 50 m thick succession with Carboniferous Culm-like facies, starting with a thin conglomerate bed at the base, followed by greenish coarse-grained micaceous metarkoses alternating with quartzose metaconglomerates and metapelites, and including one thin volcanic level.

Pizzo Leo 1 section

The Pizzo Leo 1 section (A-B in Fig. III.4.6; Fig. III.4.7) is situated along the western slope of the Favoscuro river, parallel to the Randazzo-Floresta NR 116. It roughly seems to correspond to that studied by Majesté Menjoulas *et al.* (1986). It has also provided conodont remains, but not of Late Devonian age. Instead, it has provided the oldest conodont association found, up to now, in the LTU (see section III.4.4). In the Pizzo Leo 1 section the following main lithostratigraphic units have been recognized, from bottom to top:

a) Greenish to greyish metapelites (mainly covered by soils) with a 3 m thick layer of thinly-bedded violet and greenish metamarls and light grey metalimestones at the top.

b) Light grey medium-bedded metalimestones (Fig. 13B), up to 9 m thick, with a dm-thick intercalation of dark grey metamarls.

c) Violet and greenish metamarls, around 80 m thick, with varicoloured metapelite and minor calcschist intercalations, which include a 10 m thick interval of whitish and yellowish calcschists, and several medium- (40cm) to very thick-bedded (up to 4m) horizons of grey and white metalimestone.

d) Finally, the succession ends with dark metapelites with some intercalations of metarenites, around 30 m thick.

Pizzo Leo 2 section

The Pizzo Leo 2 section (C-D in Fig. III.4.6; figs. III.4.8 and III.4.10.a) is situated along the south-western slope of the Pizzo Leo hill, between 1275 m and 1350 m of altitude. The following lithostratigraphic units have been recognized, from bottom to top:

a) Greenish to greyish (sometimes violet to pinkish) metapelites, with minor grey to greenish metarenites with at least two lenses

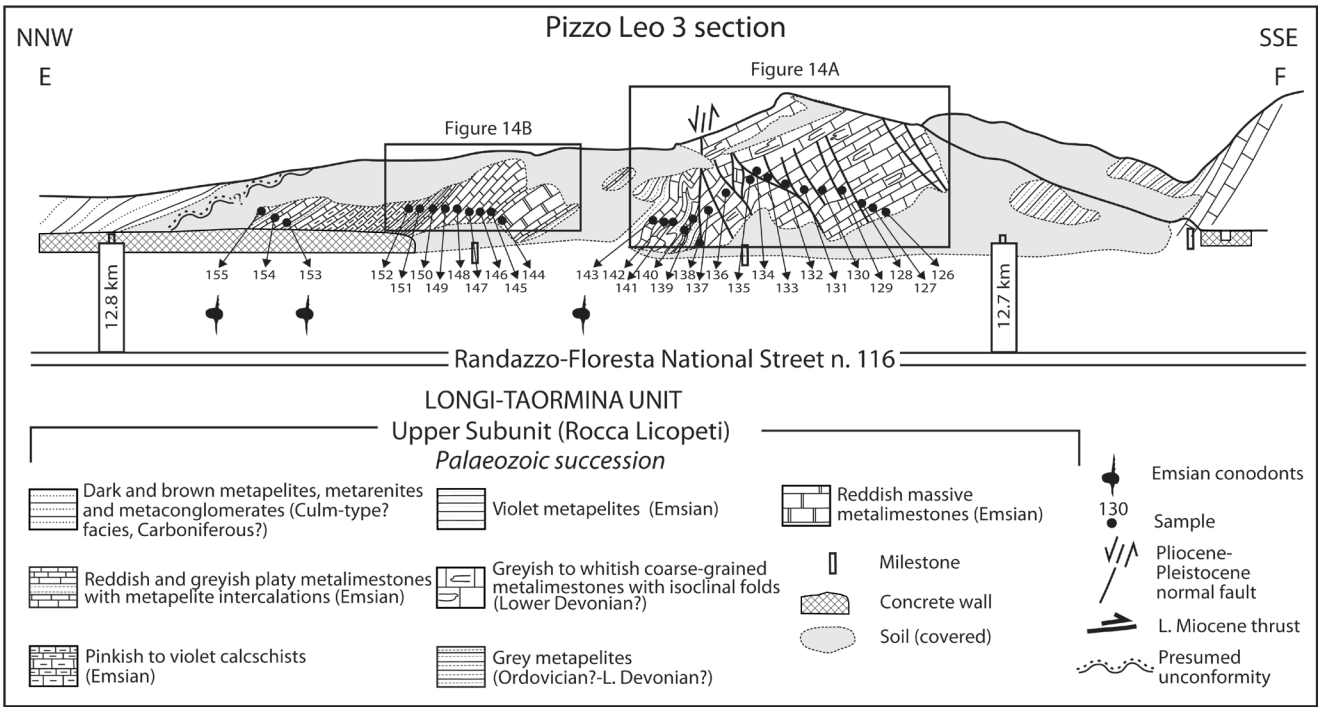


Figure III.4.9. Pizzo Leo 3 geological section (see location of the section in Fig. III.4.6).

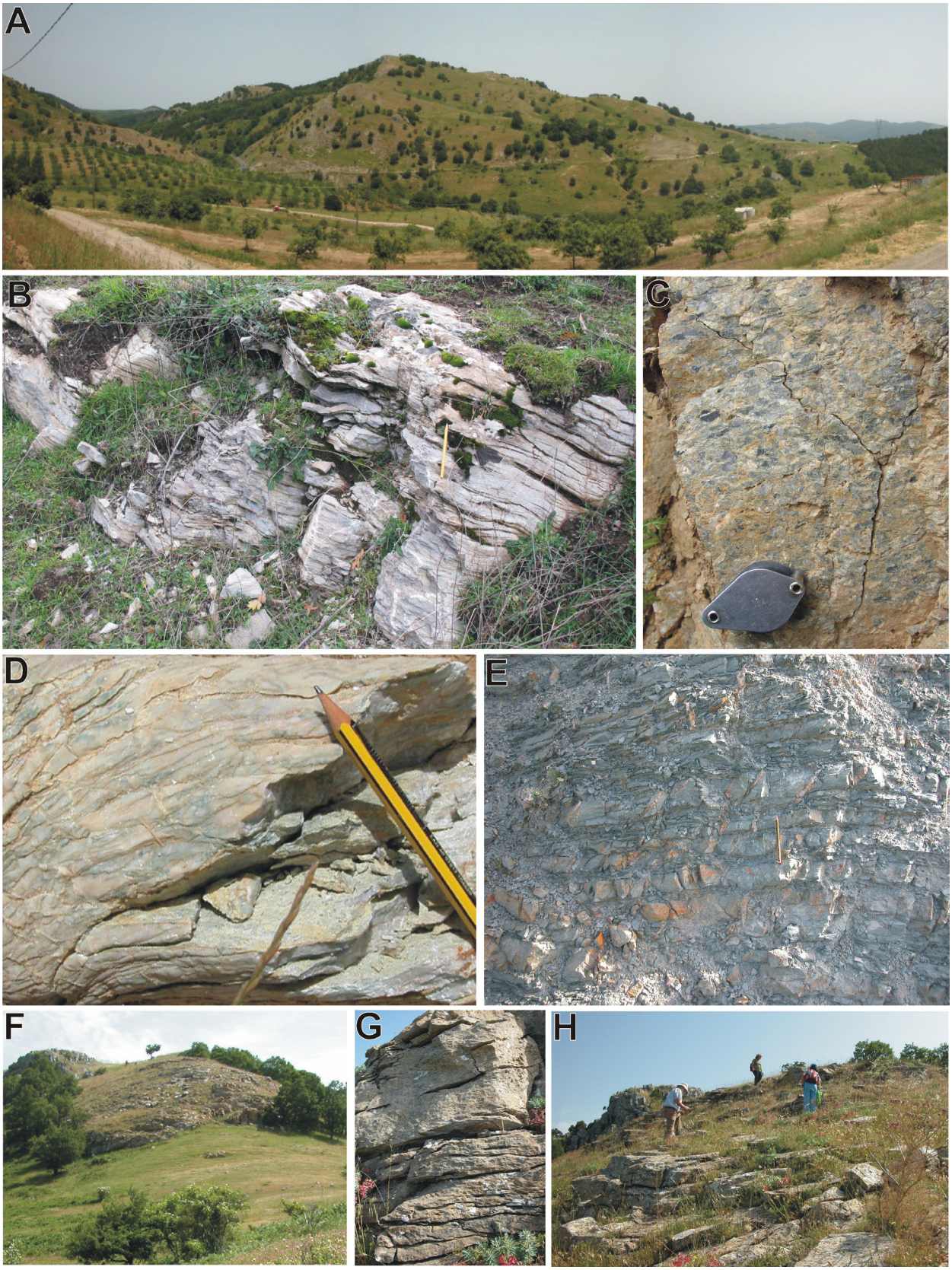


Figure III.4.10. The Pizzo Leo area. ***a,*** *Panorama of the Upper Subunit of the LTU with location of the three sections studied (south towards the right; profile at bottom approximately corresponds to that of the cross-section reproduced in Fig. III.4.8, where location of pictures (C to H) is also indicated);* ***b,*** *Silurian nodular metalimestones of the Pizzo Leo 1 section (where samples MAJ-14 and MAJ-15 were collected);* ***c,*** *Metamicroconglomerate of the lower part of the Pizzo Leo 2 section;* ***d,*** *Pinkish-whitish nodular metalimestones intercalated between metapelites in the middle part of the section;* ***e,*** *Grey metapsammites, passing upwards to grey metapelites of the upper part of the succession;* ***f,*** *Panorama of the latest Silurian-earliest Devonian metalimestones of the upper part of the section (Pizzo Leo hill);* ***g,*** *Thick to medium bedded metalimestones;* ***h,*** *Lower Devonian, well-bedded and platy metalimestones of the uppermost part of the section.*

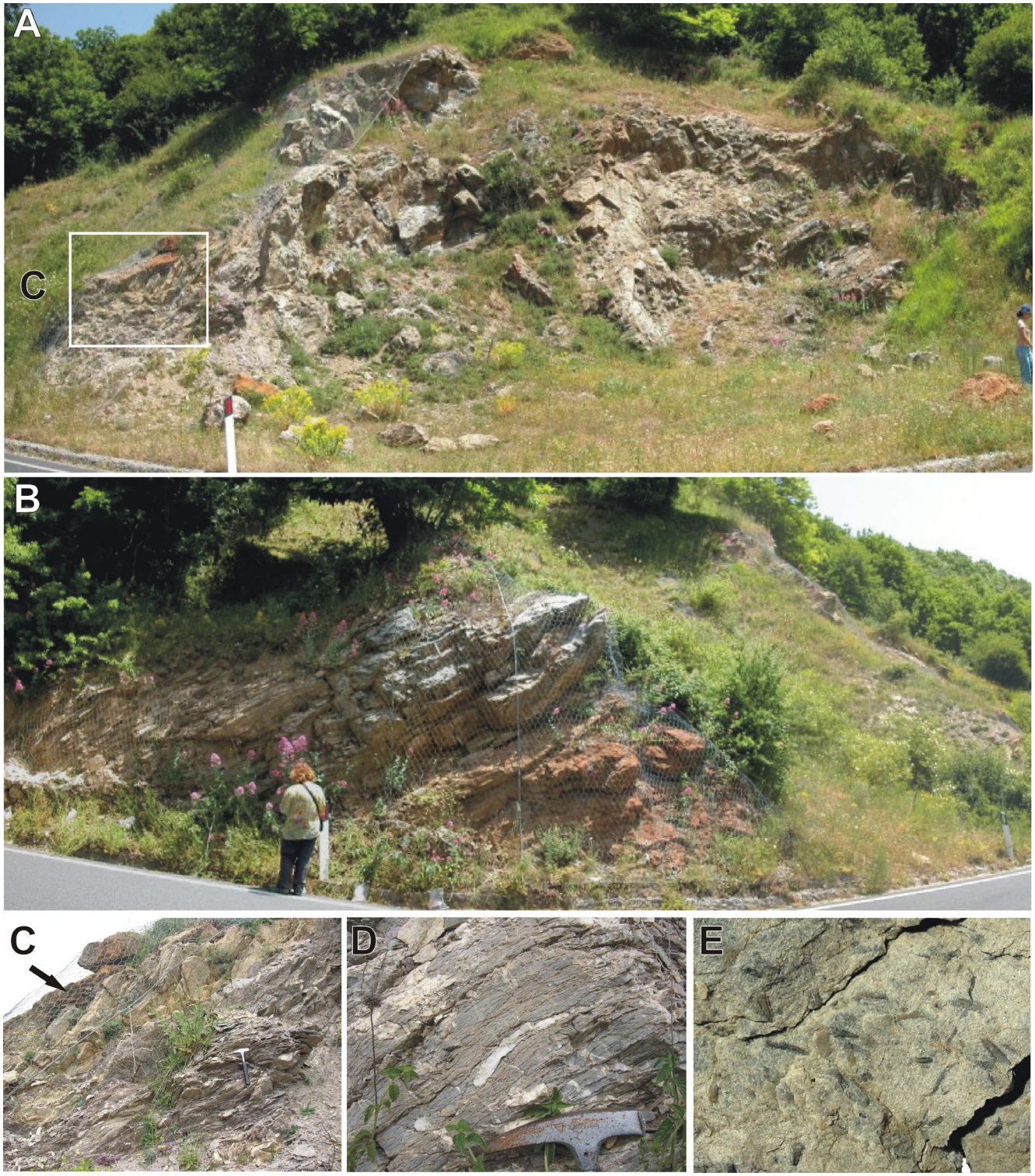


Figure III.4.11. Field views of the Pizzo Leo 3 section (see location of photos in Fig. III.4.9). ***a,*** *Massive (right) to platy (left) metalimestones of the lower part of the section;* ***b,*** *Massive and iron-rich metalimestones, passing upwards to bedded metalimestones of the upper part of the section;* ***c,*** *Close-up of the area squared in (a), corresponding to pinkish calc-schists in the transition zone from massive to well-bedded metalimestones;* ***d,*** *Strongly deformed nodular metalimestone bed;* ***e,*** *Wheathered surface of the metalimestone arrowed in (c) full of dacryoconarids (sample 07P-143).*

(up to 10 m thick) of greenish metabasalts and metadolerites in their lowermost part (Fig. III.4.8). Rare microconglomerates (Fig. III.4.10.c) and metre- to decimetre-thick carbonate lenses (whitish to pinkish, locally nodular, metalimestones(Fig. III.4.10.d), and calc-schists are intercalated in the metapelites and metarenites overlying the metavolcanites. Gradually, the succession evolves upwards to grey metapelites and metarenites (Fig. III.4.10.e), with microconglomerate intercalations and carbonate lenses in the upper part. The total thickness of metapelites is around 80 m. This interval passes gradually upwards into carbonates (Fig. III.4.10.f).

b) Greyish to beige (reddish when weathered) thick-bedded (metric), strongly deformed and recrystallized limestones, up to 20 m thick, which provide conodonts of earliest Devonian age (see section III.4.3). They are thick bedded and nodular at the base (Fig. III.4.10.g), and change upwards into medium-bedded to platy limestones (Fig. III.4.10.h).

c) The Paleozoic succession ends with around 30 m of dark and brown metapelites, metarenites and metaconglomerates (Fig. III.4.8), and it is unconformably covered by Triassic?-Hettangian continental red beds with Pseudoverrucano lithofacies.

Pizzo Leo 3 section

The Pizzo Leo 3 section (E-F in Fig. III.4.6; figs. III.4.9 and III.4.11) is located along southeastern edge of the Randazzo-Floresta 116 NR (between km 12,7 and 12,8). This succession is made of, from bottom to top:

a) Grey metapelites mainly covered by soils.

b) Greyish to whitish (reddish when weathered) thick-bedded and coarse-grained, strongly deformed and recrystallized limestones with isoclinal folds, up to 20 m thick (Fig. III.4.11.a). They change upwards into a more platy succession (Fig. III.4.11.b). The bed bearing dacryoconarids (Fig. III.4.11.e) and firstly observed by Lardeaux and Truillet (1971) is located exactly in the zone where the bed thickness changes in this section (Fig. III.4.11.c). Above it, and after a badly exposed interval of metapelites and calc-schists, a massive, iron-rich reddish metalimestone passes upwards to bedded greyish metalimestones that gradually evolve to the above mentioned more platy succession. It is up to 30 m thick, and is made of pinkish to violet calcschist with some gray platy metalimestone beds with metapelite intercalations. These rocks have provided the youngest conodonts associations found in the Pizzo Leo area (see section III.4.3).

c) Brown slates, where Majesté-Menjoulas *et al.* (1986) also found few metres of lydites and volcanites, which we have not found probably due to its reduced thickness and to recent cree and soil cover on this part of the section (Fig. III.4.9). The succession ends with dark and brown metapelites, metarenites and metamicroconglomerates.

III.4.3. Conodont biostratigraphy

Several outcrops of metalimestones have been sampled for conodonts in the eastern (Gallodoro, Lower Subunit; Castelmola, Middle Subunit; Fig. III.4.2) and central sectors (Pizzo Leo, Upper Subunit, Fig. III.4.1) of the LTU.

A total number of 67 samples, with a total amount of 59 kg, have been dissolved. Only 24 have provided conodonts remains but most of them are fragments that can hardly be identified as conodonts. Nonetheless, scarce and poorly preserved, but identifiable, conodont elements have been found in 2 of 7 samples from the Pizzo Leo 1 section (Fig. III.4.7), in 3 of 22 samples from the Pizzo Leo 2 section (Fig. III.4.8) and in 3 of 30 limestone samples from the Pizzo Leo 3 section (Fig. III.4.9). The great majority are little fragments that could not be classified even at generic level, but, among them, some determinable more or less complete conodont elements (Figs. III.4.12 to III.4.14), ranging from the late Silurian to the Early Devonian (Emsian) could be identified.

Pizzo Leo 1 section

Two samples collected in the same limestone interval (MAJ-14 and MAJ-15: figs. III.4.7 and III.4.10.b) have released determinable conodonts in the Pizzo Leo 1 section (Fig. III.4.12). Sample MAJ-14 yielded several fragments that are attributed to Pa elements of *Ancoradella ploeckensis* Walliser 1964. These fragments are not excavated in lower view and show a highly denticulated carina on sinuous processes with platform. Such features fit well with the stelliplanate Pa element of *Ancoradella ploeckensis*, which is Ludlow in age (Fig. III.4.12.1-3). This finding is important because Silurian rocks have

not yet been dated in the LTU. Further sampling and analysis are necessary in order to confirm this age. Sample MAJ-15 yielded one Sa element known in morphotaxonomy as *Trichonodella inconstans* Walliser 1957 and one Sb element of the morphotaxon (mf.) *Lonchodina* cf. *greilingi* Walliser, 1957 (Fig. III.4.12.4 and 5 respectively). In multielemental taxonomy, both elements are included as part of, among others, the apparatuses of *Oulodus elegans* (late Silurian to Early Devonian) and *Kockelella variabilis* (Late Silurian).

Pizzo Leo 2 section

In the Pizzo Leo 2 section, three samples (07P-174, 07P-175 and 07P-176: Fig. III.4.8) yielded determinable conodonts, all of them located in the medium-bedded reddish limestones of the upper part of the section (Fig. III.4.13). Sample 07P-174 released *Icriodus* cf. *woschmidti* Ziegler 1960 and *Ozarkodina remscheidensis* (Ziegler 1960). The Pa element of *Icriodus* found in this sample is fragmented and very badly preserved but the main features of *I. woschmidti* are visible (Fig. III.4.13.2 and 3), such as the presence of transverse ridges instead of three rows of nodules, a weak thin median longitudinal ridge in the grooves between the transverse ridges and a lateral process. The Pa element of *Ozarkodina remscheidensis* (Fig. III.4.13.1) is broken and twisted, but the presence of a strongly expanded and slightly asymmetrical basal cavity, and the mostly irregular denticle height, particulary the one above the basal cavity, is typical of this species. The presence of *Icriodus woschmidti* indicates a Lochkovian age for these beds.

Sample 07P-175 yielded *Ancyrodelloides omus* Murphy and Matti 1983 transitional to *Ancyrodelloides transitans* (Bischoff and Sannemann 1958) and *Pseudooneotodus beckmanni* (Bischoff and Sannemann 1958). Sample 07P-176 released *Ancryrodelloides* cf. *omus* Murphy and Matti 1983 and diverse elements of *Oulodus* sp. The recovered specimen of *Ancyrodelloides omus-transitans* is badly preserved (Fig. III.4.13.4) but it can be

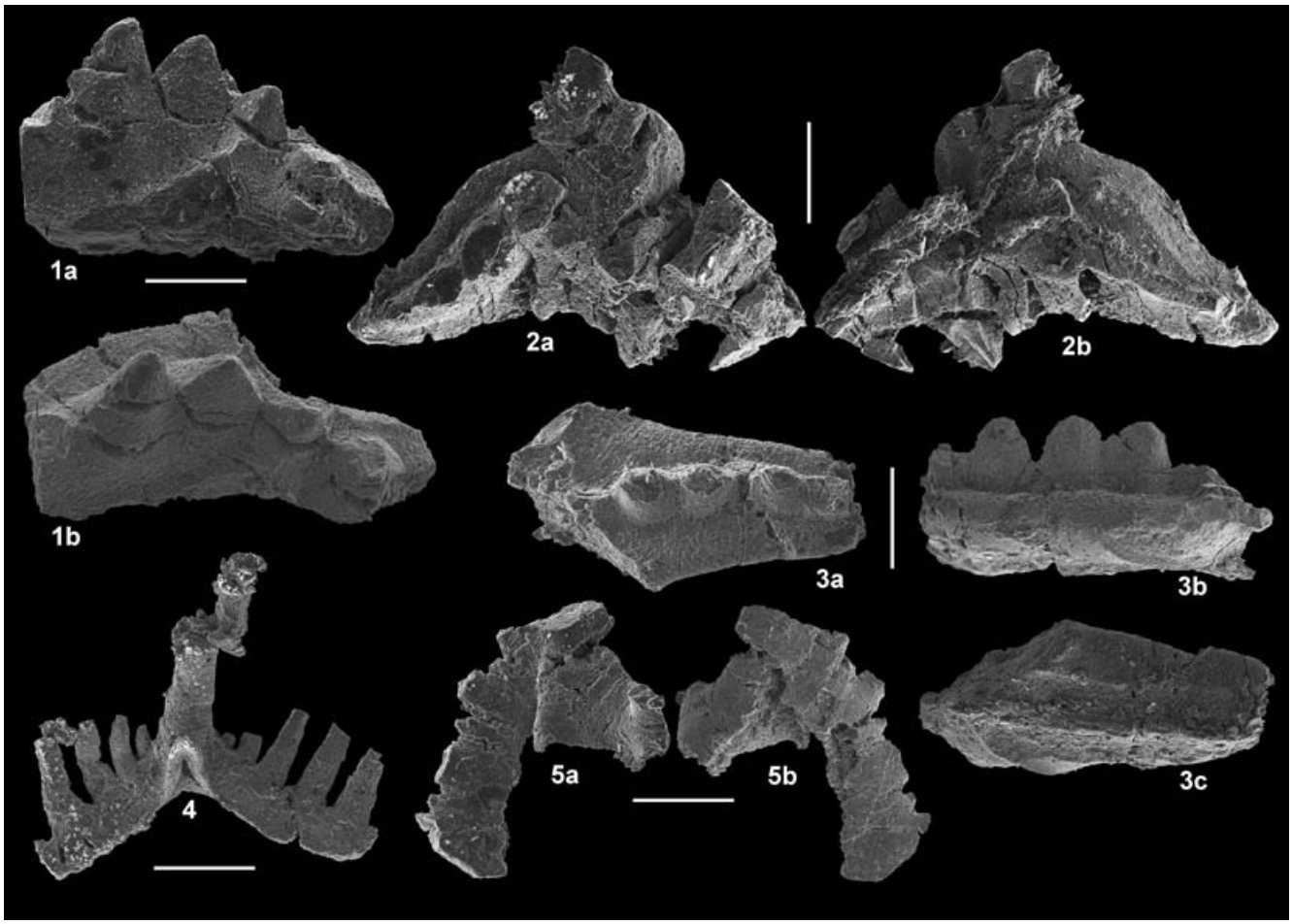


Figure III.4.12. Conodonts of the Pizzo Leo 1 section. Scale bar are 200µm.
1-3, cf. *Ancoradella ploeckensis; Pa element (fragments).* ***1a,*** *MAJ-14-2, lateral view;* ***1b,*** *MAJ-14-2,upper view;* ***2a,*** *MAJ-14-3, upper view;* ***2b,*** *MAJ-14-3, lower view;* ***3a,*** *MAJ-14-1, upper view;* ***3b,*** *MAJ-14-1, lateral view;* ***3c,*** *MAJ-14-1, lover view.* ***4,*** *Sa element (*mf. *'Trichonodella inconstans'); MAJ-15-1, posterior view.* ***5,*** *Sb element (*mf. cf. *'Lonchodina greilingi'); MAJ-15-3;* ***5a,*** *Anterior view;* ***5b,*** *Posterior view.*

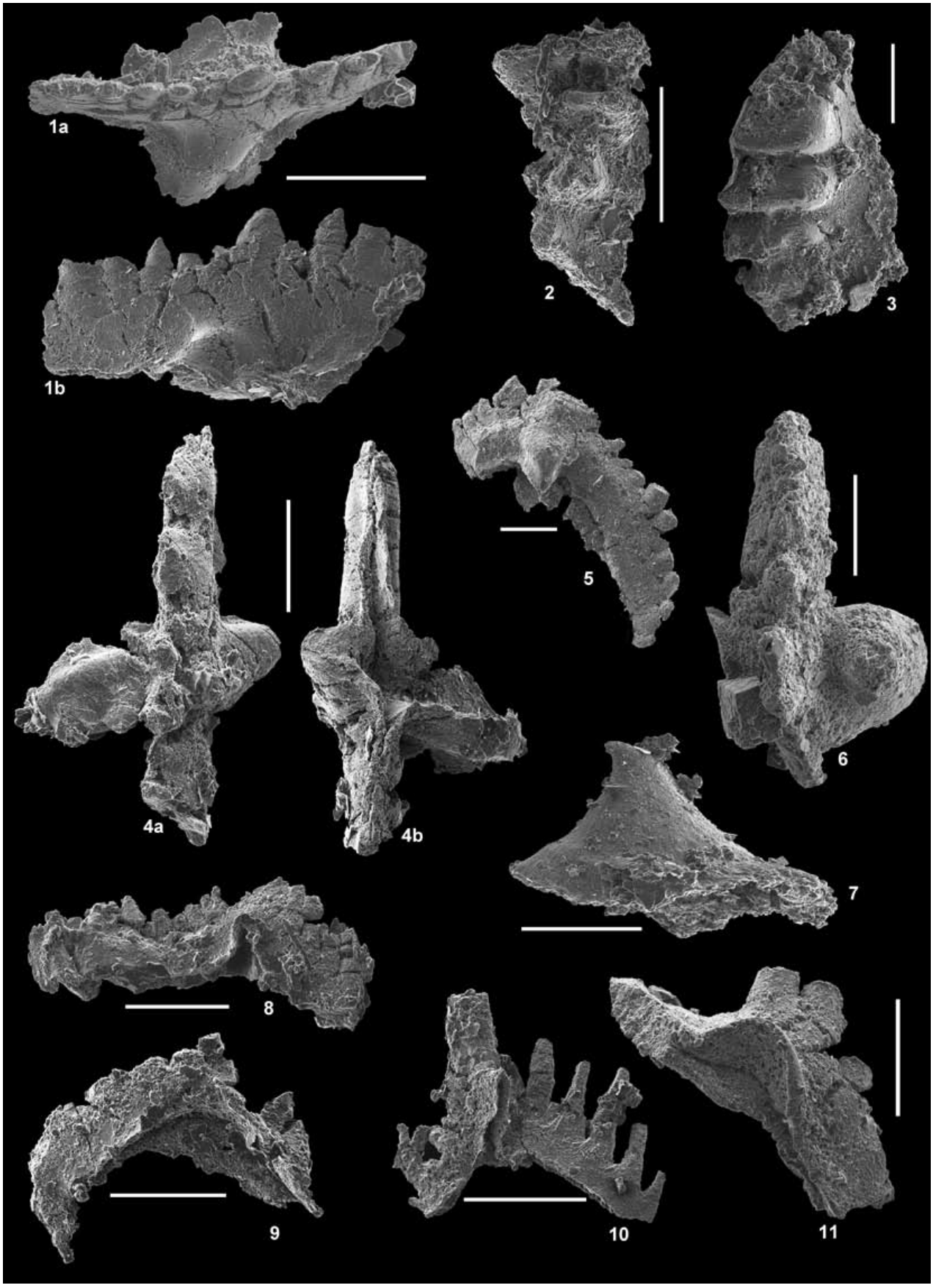


Figure III.4.13. Conodonts of the Pizzo Leo 2 section. Scale bars are 200µm.
1, *Ozarkodina remscheidensis remscheidensis; Pa element 07P-174-4.* ***1a,*** *Upper view;* ***1b,*** *Lower view.* ***2,*** *Icriodus cf. woschmidti; Pa element 07P-174-2.* ***3,*** *Icriodus cf. woschmidti; Pa element 07P-174-1.* ***4,*** *Ancyrodelloides omus transitional Ancyrodelloides transitans; Pa element 07P-175-1.* ***4a,*** *Upper view;* ***4b,*** *Lower view.* ***5,*** *Ancyrodelloides* sp.*; Pb element 07P-176-6.* ***6,*** *Ancyrodelloides omus; Pa element 07P-176-1.* ***7,*** *Pseudooneotodus beckmanni; Pa element 07P-175-3.* ***8-10,*** *Oulodus* sp.*;* ***8,*** *Sb element 07P-176-4;* ***9,*** *Pa? element 07P-176-3;* ***10,*** *Sb element 07P-176-9;* ***11,*** *Pb element 07P-176-2.*

identified by the presence of nodules in one of the the shouldered platform lobe, like *A. omus*, but the posterior processes does not bent laterally and the basal cavity is a little bit constricted, like *A. transitans* (Fig. III.4.13.5 and 6). This species belongs to the *delta* Zone (Lockhovian, Lower Devonian).

Pizzo Leo 3 section

In the Pizzo Leo 3 section, three samples (07P-143, 07P-153 and 07P-155: Fig. III.4.9) yielded datable conodont elements. The location of the lowest fossiliferous sample (07P-143) coincides with the (geometrically) lowest basal beds of the platy limestone succession and was taken from the same bed bearing dacryoconarids mentioned above (figs. III.4.9 and III.4.11.e). This conodont association includes *Icriodus* gr. *bilatericescens* Ziegler 1956, *Icriodus* aff. *beckmanni* Ziegler 1956 and *Polygnathus luciae* Martínez-Pérez and Valenzuela Ríos 2011 in Martínez-Pérez *et al.* (2011). A nearly complete Pa element of *Icriodus* is assigned to the *Icriodus bilatericrescens* group without specifying the subspecies (Fig. III.4.14.3). In spite of having a lanceolate spindle and two well-developed processes forming a 180º angle, the stage of development of these two processes is slightly different from those of the nominative subspecies. Even though the element is partially broken, it shows a very well developed posterior process, and an anterior process ornamented by a crest instead of by nodules. Other partially broken Pa element is assigned to *Icriodus* aff. *beckmanni* (Fig. III.4.14.4), since it shows all the characteristic features of this species but the angle between the posterior outer process and the cuspid is slightly smaller, a fact that could be explained by tectonic distortion, as the sampled rocks are strongly deformed. The studied specimens of *Polygnathus luciae* are characterized by an elongated platform with a characteristic "L"-shape, with a more developed outer platform in its posterior region, and by a slightly asymmetrical and inverted basal cavity (Fig. III.4.14.1 and 2). This taxon is closely related to *Polygnathus mashkovae* (Fig. III.4.14.9), but it is clearly distinguished from it by the presence of a clear constriction in the anterior part of the element prior the expansion of the platform in the later species. The presence in this sample of *Icriodus* gr. *bilatericrescens*, *Icriodus* aff. *beckmanni* just allows us to assign this level to the Emsian. However, the record of *Polygnathus luciae* allow us a more precise dating, indicating an early Emsian age, lower part of the nothoperbonus Zone (sensu Martínez-Pérez *et al*., 2011).

In the topmost part of Pizzo Leo 3 section two samples taken in the upper calcschists (07P-153 and 07P-155: Fig. III.4.9) yielded determinable conodonts. Sample 07P-153 released one Pa element of *Polygnathus* cf. *nothoperbonus* Mawson 1987. This element presents a long and narrow platform, characteristic of *Polygnathus nothoperbonus*; however, their posterior end is broken, and the initial posterior inversion of the basal cavity is not clearly visible. In addition, the typical transversal ridges that ornamented their tongue cannot be observed (Fig. III.4.14.5), although this could be explained because, probably, our specimen is a juvenile element according to its small size (around 0.5 mm in length).

The uppermost conodont association in the Pizzo Leo 3 section has been found in sample 07P-155 (Fig. III.4.9). This level has yielded the characteristic Emsian conodonts *Critheriognathus steinhornensis* (Ziegler 1956), *Polygnathus mashkovae* Bardashev 1986 and *Polygnathus nothoperbonus* Mawson 1987 (Fig. III.4.14.6). *Critheriognathus steinhornensis* is clearly identified by the even denticulation of the blade and the rounded and clearly asymmetric lobes (the outer lobe being more developed than the inner one), which are located closer to the posterior half of the unit (Fig. III.4.14.7 and 8). *Polygnathus mashkovae* has an asymmetric platform, with a narrow anterior third bearing a characteristic constriction and, at the posterior two-thirds, a broadly expanded platform developing a flange-like structure at the outer edge (Fig. III.4.14.9). In addition, the tongue, with continuous transverse ridges, is markedly deflected inwards and, in lower view, the basal cavity is clearly inverted. Finally, *Polygnathus nothoperbonus*, as commented above, has a characteristic long and narrow platform, where there is no high flange-like development of the outer edge, and a short tongue deflected inwards and bearing continuous ridges; these features allow us to discern between this species and the above mentioned (Fig. III.4.14.6). This conodont association also belongs to the *nothoperbonus* Zone. According to Bultynck (1989, for the equivalent *gronbergi* Zone) and Martínez-Pérez *et al.* (2011) the appearance of *Po. mashkovae* indicates an upper position within this zone. Consequently, the conodont association allows us to assign the calcschists also to the upper part of the early Emsian.

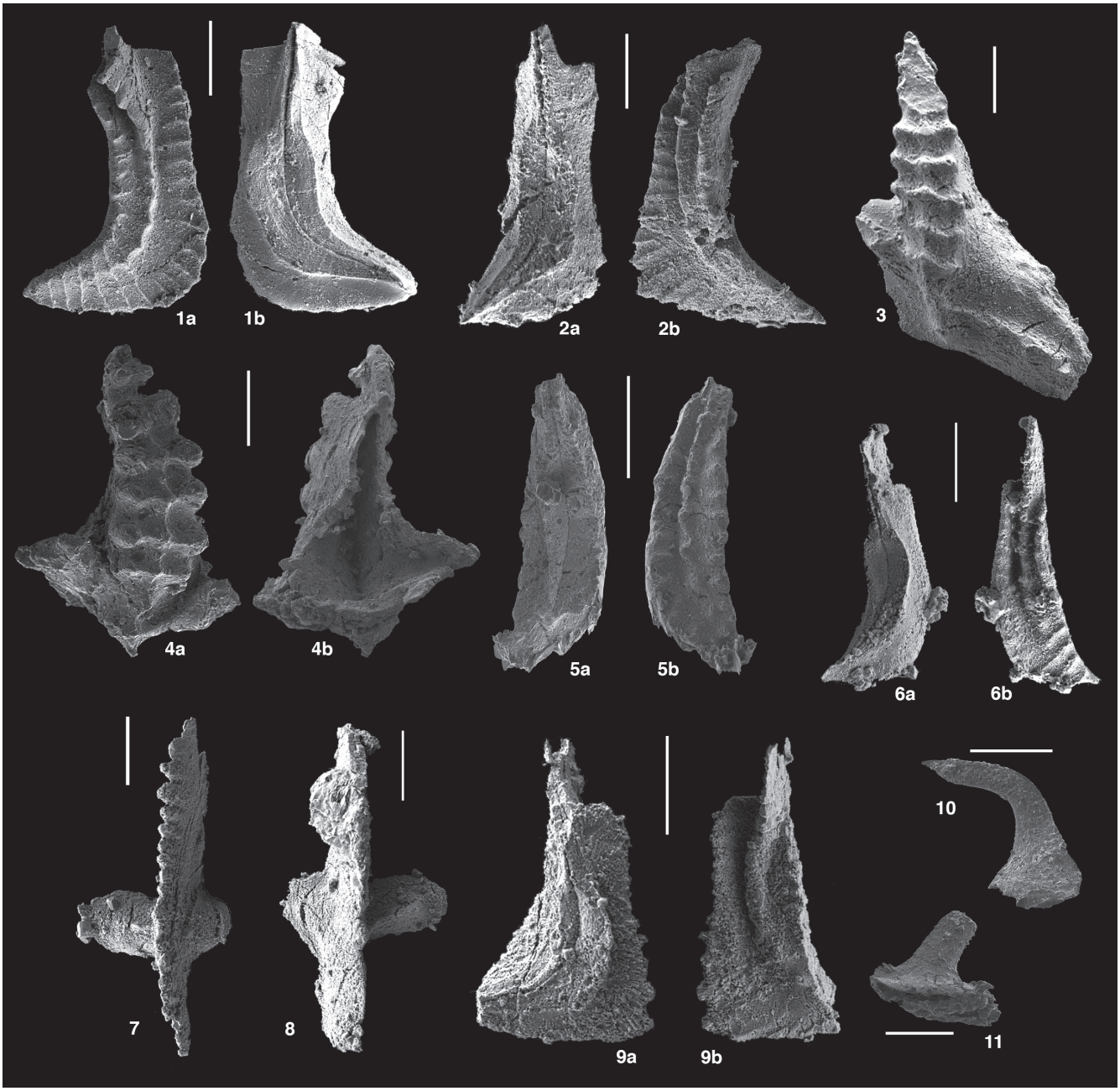


Figure III.4.14. Conodonts of the Pizzo Leo 3 section. Scale bars are 200µm.
1, *Polygnathus luciae; Pa element 07P-143-1.* ***1a,*** *Upper view;* ***1b,*** *Lower view.* ***2,*** *Polygnathus luciae; Pa element 07P-143-7.* ***2a,*** *Lower view;* ***2b,*** *Upper view.* ***3,*** *Icriodus* gr. *bilatericrescens; Pa element 07P-143-2.* ***4,*** *Icriodus aff. beckmanni; Pa element 07P-143-9.* ***4a,*** *Upper view;* ***4b,*** *Lower view.* ***5,*** *Polygnathus* cf. *nothoperbonus; Pa element 07P-153-1.* ***5a,*** *Lower view;* ***5b,*** *Upper view.* ***6,*** *Polygnathus nothoperbonus; Pa element 07P-155-1.* ***7,*** *Critheriognathus steinhornensis; Pa element 07P-155-7.* ***8,*** *Critheriognathus steinhornensis; Pa element 07P-155-6.* ***9,*** *Polygnathus mashkovae; Pa element 07P-155-4.* ***9a,*** *Lower view;* ***9b,*** *Upper view. 10-11. Icriodus* sp. *Sa elements;* ***10,*** *07P-143-12.* ***11****, 07P-143-17.*

III.5. Paleozoic tectono-stratigraphic evolution of the Longi-Taormina Unit

III.5.1. Introduction

The stratigraphic synthesis (Fig.III.4.3) and age updating presented above allow a better characterization of the Paleozoic history of the LTU. It also makes possible: i) to correlate the LTU with other better known Paleozoic successions of the Calabria-Peloritani Terrane and of nearby Alpine and Variscan areas around the Western Mediterranean, ii) to recognise lateral changes in the stratigraphic successions, iii) to interpret their sedimentary significance, and iv) to improve the picture on the palaeogeographic evolution of these (up to now poorly known) successions. Nonetheless, the reconstruction of the Paleozoic palaeogeography and plate configuration of the Paleozoic Intra-Alpine terrains (in the sense of Stampfli, 1996) is a difficult task, due to the superimposed pre-Alpine and Alpine tectono-metamorphic histories. Most available models on Paleozoic palaeogeography and plate configuration of Southern Europe and North Africa practically take no notice of any recent available geologic information (stratigraphic, in particular) coming from the Intra-Alpine Paleozoic terrains of the Western Mediterranean Alpine belts, apart from that from the Alps. The surface extent of these terrains is quite important, however, and the information derived from them must be incorporated to test the validity of such models.

The importance of these Intra-Alpine terrains lies in their key position in the Northern Gondwana margin during the opening and closure of the Paleotethys: they were located to the east of the European Variscan domains included in the Armorica composite terrane, from which most of Variscan Iberia and Corsica-Sardinia made part. Some models, mainly based on plate tectonic reconstructions, interpret that the Paleotethys was continuous along the northern border of Gondwana, from southern Asia to northern South-America, and that both the European Variscan and or the Intra-Alpine Paleozoic terrains made part of the same (European Hunic, or Galatian) composite superterrane (Stampfli and Borel, 2002, 2004; Von Raumer and Stampfli, 2008). Other models, mainly based on field studies, affirm that Paleotethyan sutures are absent in the Variscan massifs of Spain and Morocco, and actually do not ascribe any of the several ophiolite-bearing sutures present in these areas to the Paleotethys but to the Rheic Ocean (Arenas *et al.,* 2007, 2009; Simancas *et al.*, 2005, 2006; Martínez-Catalán *et al.,* 2009; Keppie *et al.,* 2010; Michard *et al.*, 2010; Murphy *et al*., 2009, 2010; Nance *et al.*, 2010; Ribeiro *et al.*, 2010; Braid *et al*., 2011), to smaller oceans like embayments related to the Rheic opening (e.g., the South-Armorican Ocean concept of Paris and Robardet, 1990) or to back-arc rifting (Azor *et al.*, 2008; Von Raumer *et al.*, 2008; Simancas *et al*., 2009) or, finally, to much older (Proterozoic) oceans (Murphy *et al*., 2006; El Hadi et al., 2010). Within the proper Alps there are ophiolitic remnants (Loth *et al.,* 2001; Schaltegger et al., 2003) that are not interpreted as related to the Palaeotethys, but as pieces of some of these smaller oceas (e.g., the Chamrouse Ocean of Von Raumer and Stampfli, 2008). In any case, these basins opened to the N and to the W of the LTU and related Western Mediterranean Paleozoic Intra-Alpine terrains, whereas the true Paleotethyan basins opened toward the E of them. In this frame, the Calabria-Peloritani Palaezoic successions should be located closer to the Paleotethys than other Western Mediterranean Intra-Alpine terrains.

One key to evaluate the palaeogeographic relations of the LTU successions with the above mentioned large oceanic areas and associated margins is to check its correlation and lateral facies evolution to neighbour domains. The history of the LTU started with deposition of Cambro-Ordovician marine fine-grained siliciclastics. In Late Ordovician time, this sedimentation was associated with an acidic volcanism related to a postcollisional rifting (Trombetta *et al*., 2004), which was followed by an alkaline basic volcanism, typical of rifted intraplate

settings (Ferla, 1978; 2000; Ferla and Azzaro, 1978; Atzori et al., 2001; Cirrincione et al., 2005) and by Devonian-Carboniferous basinal carbonates, radiolarites and siliciclastics. This evolution records the opening of a continental margin basin. In addition, several stratigraphic and volcanic features of the LTU Paleozoic can be compared to those of the Calabrian Stilo and Sila Units (Bouillin et al. 1987; Navas-Parejo *et al*., 2009a and b), of the Malaguide-Ghomaride Complex of the Betic Cordillera and the Rif (Chalouan, 1986; Chalouan and Michard, 1990; Rodríguez-Cañero, 1993; O'Dogherty *et al.,* 2000; Martín-Algarra *et a*l. 2000, 2004, 2009 a and b; Navas-Parejo et al., 2008; Rodríguez-Cañero *et al*., 2010), and, especially, of the Alps (Neubauer *et al*., 1994; Neubauer and Handler, 2000; Schönlaub and Histon, 1999, 2000; Schönlaub, 2000; Vai, 2001), which also record typical continental margin evolutions.

III.5.2. Cambro-Ordovician palaeogeographic and geodynamic scenario of the LTU

The weakly metamorphic pelites bearing Cambrian-Ordovician acritarchs of the LTU are equivalent to similar rocks that occur in the Sila Massif in Northern Calabria (Bouillin *et al*., 1984), and in the Stilo Unit. The latter, in particular, also includes metabasite bodies within a metasiliciclastic succession underlying well dated Devonian beds (Navas-Parejo *et al.,* 2009a). The lowermost metapelitic succession of the LTU also reminds the roughly coeval pelitic-psammitic successions included in the Morales Formation of the Malaguide Complex (Martín-Algarra *et al*., 2004, 2009a) and, especially, those of the Val Visdende Group of the Carnic Alps and equivalent Austroalpine formations (Piller *et al*., 2004). The latter are made up of shales, slates, laminated siltstones, sandstones, arkoses, quartzites and greywackes that, like in the LTU, are associated with intermediate to acidic porphyroids of Late Ordovician age (Comelico Porphyroid and Fleons Formation; see Schönlaub and Histon, 1999 and references therein). Consequently, as proposed for the Alpine region (Stampfli, 1996; Stampfli and Borel, 2002, 2004; Schönlaub and Histon, 1999) and also pointed out by Trombetta *et al.* (2004), the deposition of the clastic Cambrian-Ordovician succession of the LTU took place in marine environments connected to calc-alkaline volcanic arc systems related to subduction.

The Ordovician intermediate to acidic rocks of the LTU prompt the magmatic (both plutonic and volcanic) Cambro-Ordovician rocks that occur in the Iberian Massif, the Pyrenees and Sardinia (Montero and Floor, 2004; Helbing and Tiepolo, 2005; Valverde-Vaquero *et al*., 2005; Giacomini *et al*., 2006; Bea *et al*., 2007; Castiñeiras *et al*., 2008; Montero *et al*., 2009: Oggiano et al., 2010; Liesa *et al*., 2011, and references therein) and also in the Calabro-Peloritanian Aspromonte Unit (Fornelli et al., 2007), although the age of the equivalent magmatism in the LTU is slightly younger, in general (Trombetta *et al*., 2004). These features seems to relate the evolution of the LTU to that of Variscan regions of Iberia. Like them, the LTU was part, during the Ordovician, of the hanging-wall block of an active continental margin. This geodynamic scenario was probably related to the post-Cadomian to Cambrian-Ordovician subduction of the Prototethys Ocean under Northern Gondwana (Fig. III.5.2). According to most authors working in the Alps, the subduction zone was located northwards (in present-day coordinates) of Gondwana-derived Intra-Alpine microcontinents during the Latest Proterozoic and Earliest Paleozoic (Eichorn *et al*., 2001; Neubauer, 2002; Stampfli and Borel, 2002, 2004; Stampfli *et al*., 2002; Von Raumer *et al*., 2003; Schulz *et al*., 2004; Von Raumer and Stampfli, 2008). This Pacific-type, long-lasting (Neoproterozoic to Early Paleozoic) subduction generated a widespread intra-arc or back-arc Cambro-Ordovivian calc-alkaline to alkaline acid to intermediate magmatism everywhere along Northern Gondwana (see Gutiérrez-Alonso *et al*., 2003; Linnemann *et al.,* 2008; Díez Fernández *et al*., 2012; Williams *et al*., 2012, among others), and it predates postcollisional docking and amalgamation with other terrains related to the Armorica composite terranes located northwards of the Alpine region (Von Raumer, 1998; Stampfli and Borel, 2004; Von Raumer and Stampfli, 2008). In the LTU, however, as well as in some sectors of Sardinia (Oggiano et al., 2010), the calc-alkaline volcanism associated with the Cambro-Ordovician rifting ended a little bit later than in most European Variscan regions.

III.5.3. Age and geodynamic-palaeogeographic significance of the alkaline mafic volcanism

Before the present study, Silurian rocks had not been dated in the LTU. The conodont-bearing, Uppermost Silurian-Lowermost Devonian nodular and platy limestones found in the Pizzo Leo area on top of the metapelites bearing metabasite intercalations, indicate that the rifting-related marine siliciclastic sedimentation in the LTU followed with similar features during the Silurian, but with changing volcanic conditions: from Ordovician calc-alkaline acidic-intermediate volcanites to younger alkaline basic vulcanites,. Consequently, although a Cambrian-Ordovician age was proposed by some previous studies for the alkaline volcanism of the LTU, stratigraphic data definitely point to a mainly Silurian age for the main episode of basic vulcanism (maybe also Devonian *pro parte*, in some areas where Devonian beds lie systematically above the alkaline metabasites). New isotopic data from the metabasites themselves would be needed to confirm this age, however.

The widespread occurrence of alkaline metabasalts and related metavolcanoclastic rock in all outcrops of the LTU evidences the progress of the intracontinental rifting in Silurian-Devonian time. This evolutionary frame resembles that proposed for the opening of the Paleotethys in many alpine areas (Stampfli and Borel, 2004) and indicates that the LTU was involved in rifting events related to the early (Silurian) evolution of the Paleotethys. Alkaline metabasalts actually predominate over pelitic rocks in wide areas of the LTU, especially in its westernmost outcrops, which were probably located in a sector of the basin that was strongly affected by extensional tectonics. If the important late Alpine clockwise rotation that affected the LTU (Somma, 2006) and the anticlockwise rotations in which the Calabrian sector of the orogen was involved are taken into account, the western outcrops of the LTU, where the Paleozoic successions are dominated by basic volcanics, should be originally located in a southernmost (perhaps south-easternmost) position than all remaining units of the Calabria-Peloritani Terrane. According to the generaly accepted palaeogegraphic scenario, such area should be closer than other Paleozoic Intra-Alpine terrains to the oceanic domains of the Paleotethys, opened towards the East in Devonian-Carboniferous time.

Basic vulcanites are also present in other units of the Calabria-Peloritani Terrane and in equivalent Betic and Maghrebian Intra-Alpine terrains located towards the west, but their volume is much lesser than in the LTU. In particular, the Paleozoic succession of the Stilo Unit (Calabria) is also characterized by low-grade siliciclastic metasediments and metabasites (Pardalà schists and associated green amphibolites) that systematically underlie well dated Devonian beds (de Capoa-Bonardi, 1970; Bouillin *et al.,* 1987; Navas-Parejo *et al*., 2009a and 2009b). This means that: i) as in the LTU, the metabasites in the Stilo Unit come probably also from Silurian basic volcanics (or subvolcanics); ii) the depositional area of Stilo Unit was farer from the main volcanic centers and from the main axis of tectonic extension than the LTU coeval depositional area; and iii) the palaeogeographic position of the Stilo Unit was more marginal that that of the LTU, which was located southwards and in a distal position within the continental margin, then closer to the basin axis. The latter point is actually confirmed by facies comparison (see below).

Nevertheless, neither the geochemical features of the basic volcanics of the LTU nor the later tectono-metamorphic Variscan evolution that mildly affected this unit, evidence the formation of true oceanic crust, or its later subduction. The same accounts in other units of the Calabria-Peloritani Terrane or in equivalent domains of the Western Mediterranean Alpine Orogen in North Africa and South Spain, with the only remarkable exception of the Mela Unit of the Peloritani Mountains (Messina *et al.*, 1997), which includes Variscan or Eo-Variscan eclogites (Messina *et al.,* 2004 Macaione *et al.,* 2010). The Mela Unit, like probable equivalent rocks of Sardinia (Giacomini *et al*., 2005; Franceschelli *et al.*, 2007) and of the Alps (Von Raumer, 1998; Loth *et al.*, 2001; Schaltegger *et al.,* 2003; Stampfli and Borel, 2004, or Von Raumer and Stampfli, 2008, and references therein), should record the pre-Variscan opening of some small and northern oceanic areas probably located to the N of the LTU, as well as its pre-Variscan or eo-Variscan subduction. But apart from this, the LTU and laterally related successions in Western Mediterranean belts, from Italy to the Gibraltar Arc, were deposited in passive continental margin successions since the Latest Ordovician to the Earliest Carboniferous.

To conclude, the great development of intracontinental alkaline mafic vulcanites in the LTU must be interpreted as a marker of its eastern location within the collage of Western Mediterranean Intra-Alpine terrains, in a palaeogeographic position

closer to the Palaeotethys spreading centers than other Western Mediterranean Intra-Alpine terrains. The Silurian and younger succession of the LTU was deposited in deeper areas and in more distal region of the margin than the coeval succession of the Stilo Unit and equivalent terrains in Western Mediterranean Alpine Belts. Consequently, the LTU was palaeogeographically located on a southern or south-easternmost position relative to the above mentioned laterally related units. Taking into account that there is no geological evidence of the westward prolongation of Paleotethyan sutures in the Variscan domains of Iberia and North Africa, the Western Mediterranean Intra-Alpine terrains *in toto*, and the LTU in particular, must be interpreted as continental margin representatives of the the western termination of the Paleotethys. They should be the located in a central position of the southern margin of the European Hunic (Galatian) Terranes, in a region that should not have been completely detached from Gondwana, then close to the Northern Gondwana passive margin, and situated between the Amorica composite terrains (to the SW) and the Paleozoic domains of the Alps (to the NE.

III.5.4. Correlation, evolution and significance of the LTU sedimentary facies

The presence of Latest Silurian conodonts on top of the metapelitic successions with metabasites indicates that, during most of Silurian time, deposition in the LTU was dominated by siliciclastic sediments. Apart from the volcaniclastic contribution related to alkaline mafic volcanics, these siliciclastic sediments would not very different to those bearing acritachs and previously deposited during the Cambrian-Ordovician. This siciliclastic sedimentation was essentially fine-grained, but the local presence of quartzitic metasandstones and stretched microconglomerates indicates the action of bottom currents.

The upwards change from a siliciclastic towards a mixed siliciclastic and carbonatic deposition with conodont-bearing Silurian-Devonian pelagic limestone horizons and, at least locally, to Carboniferous siliceous sediments bearing radiolaria, indicates a Silurian to Carboniferous deepening upwards evolution of the marine environments. After intracontinental rifting under extensional regime, which is considered to be the geodynamic frame under which the alkaline volcanites of the LTU were formed, the persistent thermal subsidence that is typical of the mature evolutionary stages of divergent continental margins, reasonably explains the above mentioned transgression and deeping.

The rapid northward palaeolatitudinal drift of Gonwanaland during the Silurian (Tornsvik and Cocks, 2004; Cocks and Tornsvik, 2006, among many others) also accounts for the progressive sedimentary change from siliciclastic to carbonate sediments in the LTU. The siliciclastic sediments were probably deposited in cool to temperate seas, as most European and North African realms were located at high latitudes near to the South Pole in Ordovician time. On the contrary, the typical Silurian carbonates, frequently showing nodular facies, bearing orthoceratids, tintinnids, conodont and other fossils, were deposited in much warmer waters.This paleolatitudinal migration and sedimentary change is well documented in the Hirnantian to Silurian record of Variscan and Alpine terrains around the Western Mediterranean. In particular, the Hirnantian conodont associations recently found by Rodríguez-Cañero *et al*, (2010) in the Malaguide Complex of the Betic Cordillera are typical of cool but not glacial waters; actually, they are very different to those of coeval Upper Ordovician sediments of the Iberian Variscan regions (Sarmiento *et al.,* 2011), which were deposited in glaciomarine settings related to the Hirnantian glaciation (Gutiérrez-Marco et al., 2010). Moreover, the Hirnantian Malaguide conodont associations and sediments are much more similar to those found in the Carnic Alps than to those of the Iberian Massif, thus pointing to a palaeogeographic location of the Malaguide Complex (which is very similar to the Stilo Unit) all Western Mediterranean Intra-Alpine terrains in areas closer to the Alps than to the Variscan domains of Iberia, as also previously proposed for the LTU, then at lower palaeolatitudes than their Iberian counterparts.

The thin pinkish to whitish nodular metalimestone horizons that are locally included in the lower part of the greenish metapelites with metabasite lenses of the Pizzo Leo 2 section, as well as the sometimes nodular, conodont-bearing

Uppermost Silurian and Lowermost Devonian limestones of the Pizzo Leo 1 and 2 sections are here considered to be, according to its stratigraphic position and lithofacies, possible equivalents to the Malaguide and Carnic Alpine Uppermost Silurian carbonate lithofacies. In the Carnic Alps, the Silurian nautiloid-bearing, nodular and condensed pelagic limestones are indicative of high bottom areas of the continental margin with reduced sedimentary rate, and they change laterally either to shallower marine bioclastic limestones interbedded with shales, or to graptolite-bearing black shales (ampelites) and cherts deposited in anoxic environments or, basinwards, to fine-grained siliciclastic turbiditic successions (Schönlaub, 1998; Schönlaub and Histon, 1999; Piller *et al.*, 2004; Ferretti, 2005; Brett *et al*., 2009). This change from condensed, preferentially carbonatic pelagic successions passing laterally to pelitic sequences, represents a transgression, which homogenized the basin during the Pridoli epoch (Schönlaub and Histon, 1999).

Although typical Silurian anoxic facies have not been clearly identified in the metapelites of the LTU, they probably represent deepest marine terrigenous counterparts of the quite probably anoxic deposits (in origin) that appear in other tectonic units of the Calabria-Peloritani Terrane, e.g., the Fondachelli, Mandanici and Stilo Units. The Fondachelli and Mandanici Units tectonically overlie the LTU and are characterized, among other lithofacies, by black (graphite-rich) fine-grained low-grade metapelites whose protolith, in the Antillo area, was considered by Bouillin et al. (1987) to correspond to Silurian ampelites (black shales); unfortunately, our sampling for conodont of metalimestone lenses associated with the black metapelites in the Antillo area was unsuccessful. In the Stilo Unit (Calabria), Silurian beds have neither been properly dated, but its succession includes typical black shale horizons near the top of the metabasite-bearing metapelites (Bouillin et al., 1987), and immediately below Lower Devonian beds well dated by conodonts (Navas-Parejo *et al.*, 2009a and 2009b). This points to a lateral facies zoning in the Calabria-Peloritani Paleozoic successions, which is equivalent to that inferred above from the vulcano-sedimentary successions and which is even better evidenced when Devonian beds are correlated.

The Devonian of the Austrian Alps in the Graz area is particularly characterized by thick carbonate deposits which locally have formed build-ups (Hubmann *et al.*, 2006; Hubmann and Suttner, 2007). Laterally, to the Carnic Alps, these shallow water limestones grade into: i) pelagic carbonate sequences with reduced thickness, bearing orthoceratids, dacryoconarids and conodonts; to ii) pelitic successions sometimes bearing basic vulcanites; and iii) siliciclastic sediments, usually fine-grained, sometimes turbiditic (Schönlaub and Histon, 1999, 2000; Piller *et al.*, 2004). Although Devonian shallow marine limestones have not been recognized in any unit of the Calabria-Peloritani Terrane, the laterally related Devonian pelagic facies of the Alps are equivalent to those found in the Devonian of the Longi-Taormina and of the Stilo Units. Similar Silurian-Devonian successions can also be found in the Malaguide and Ghomaride Units around the Gibraltar Arc where local evidences of Devonian-Carboniferous carbonate platforms are also present, but as resedimented (sometimes very large) clasts included in Carboniferous conglomerates (Herbig, 1984; Chalouan, 1986; Martín-Algarra *et al*., 2004, 2009a 2009b).

In the LTU the facies associations of the Devonian limestones are poorer in fossils and richer in terrigenous intercalations than in the Stilo Unit. The carbonate facies of the Stilo Unit are thicker and richer in Devonian fossils, dacryoconariids and conodonts, the latter being particularly abundant in Upper Devonian beds (Navas-Parejo et al., 2009b). In addition, the Devonian imestones of the Stilo Units are frequently reddish, nodular and condensed, and bear orthoceratid and isolated corals in Lower Devonian beds (Navas-Parejo *et al.*, 2009a). These features indicate a depositional highbottom area within a rifted Devonian continental margin. On the contrary, the coeval grey metalimestones and grey-pinkish metamarls with greenish to greenish-violet metapelites that characterize the pre Emsian succession of the LTU in the Pizzo Leo sections here studied definitely point to deeper pelagic environments than those of the Stilo Unit. This agrees with a more distal location of the LTU within the former basin and continental margin.

The deposition of the yet undated but probably Lower Carboniferous lydites in the LTU succession fits well with the evolution found in the same age in many equivalent successions of other Alpine-Mediterranean areas, in particular in the Stilo Unit (Navas-Parejo *et al.,* 2009b) and in the Malaguide Complex of the Betic Cordillera (O'Dogherty et al., 2000). Such facies evolution points out to deposition in a deep basin below the calcite compensation depth, and indicates the maximum opening of the Calabro-Peloritanian and equivalent successions.

III.5.5. Palaeogeographic and geodynamic evolution: proposal of a new model

As regards to the geodynamic setting during sedimentation, the Silurian-Devonian tectono-stratigraphic evolution and facies pattern imply spatially and temporaryly enhanced rates of subsidence in extensional regimes. Thus the stratigraphic sequence evolves towards a mature passive margin that was related to the opening of a new ocean, the Paleotethys, after the detachment of a ribbon continent from which most Paleozoic Alpine terrains come from (Stampfli, 1996, 2000; Von Raumer, 1998; Stampfli and Borel, 2002; Stampfli *et al.*, 2002; Von Raumer *et al.*, 2002). The same is evidenced, as previously discussed, by the widespread Silurian(-Devonian?) rifting-related alkaline mafic volcanism of the LTU, which is also locally present in the Alps (Schönlaub and Heinisch, 1994) and, in a much lesser extent, in the Malaguide-Ghomaride Units of the Gibraltar Arc, where it can locally reach, as in the LTU (Acquafreda *et al*., 1991), the base of the Carboniferous (Herbig, 1993; Chalouan, 1986; Martín-Algarra *et al.*, 2009 a).

The proposed palaeogeographic model is represented in Figs. III.5.1 and III.5.2. This model assumes, according to recent zircon data of Bea *et al.* (2010), that a great part of the Gondwana-related regions of Central Iberia were not attached to Morocco in NW Africa but to the East African Orogen close to Egypt, and that the Western Mediterranean Intra-Alpine terrains were located farther to the East (Fig. III.5.1) at lower latitudes than Variscan regions of Iberia during the Early Paleozoic, according to Rodríguez-Cañero *et al.,* 2010). After Cambrian-Ordovician orogenic events related to the southward-directed Prototethyan subduction under the central sector of Northern Gondwana, a calc-alkaline magmatic suite developed in the LTU. The volcanic arc, equivalent to other Mid- to Late-Ordovician arcs formed either towards the NE, in the future Alps, or towards the W, in the Variscan regions of Iberia and Sardinia, finally collapsed due to the progress of extensional tectonics in the supra-subduction area. This determined the opening of a continental margin in the Western Mediterranean Intra-Alpine regions during the Silurian-Devonian. This margin opened towards the SE, and the LTU represent its most distal region, which was an extremely thinned intracontinental area, close to the south-westernmost end of the Paleotethys. All Western Mediterranean Intra-Alpine terrains made part of this south-western Paleotethyan embayment, and would constitute the westernmost part of the southern palaeomargin of the European Hunian (or Galatian) ribbon continent that was being detached from Northern Gondwana during the Silurian opening of the Paleotethys. Consequently, the margin formed by the Western Mediterranean Intra-Alpine terrains was faced to the new passive margin of Northern Gondwana in Silurian-Devonian time, and the LTU successions constituted the most distal region of the westernmost European Hunian (Galatian) Palaeomargin, close to the end of the Paleotethyan oceanic crust.

Finally, as it also accounts in other equivalent Alpine areas, the end of the Early Carboniferous is characterized in the LTU by deposition of siliciclastic immature deposits that were transformed into metapelites, metasandstones and metaconglomerates during the later Variscan tectono-metamorphic and magmatic evolution (De Gregorio *et al*., 2003, Fiannaca *et al.*, 2008; Appel *et al.*, 2011). These siliciclastic lithofacies can be interpreted as the typical Carboniferous flysch known as Culm facies. This indicates that the LTU succession was involved within an active plate margin during the early stages of the closure of the Paleotethys (post-Visean Carboniferous). The progress of the Variscan (Hercynian) orogenic evolution that affected the Alpine Paleozoic terrains culminated with the formation of Pangea. Nonetheless, in spite of the strong tectonic deformation and low-grade metamorphism that affects the LTU succession, many stratigraphic features and some fossils are yet preserved. This suggest that the LTU was not subducted to great depth, and that it remained in relatively shallow position within the accretionary prisms up to the end of the Variscan collision.

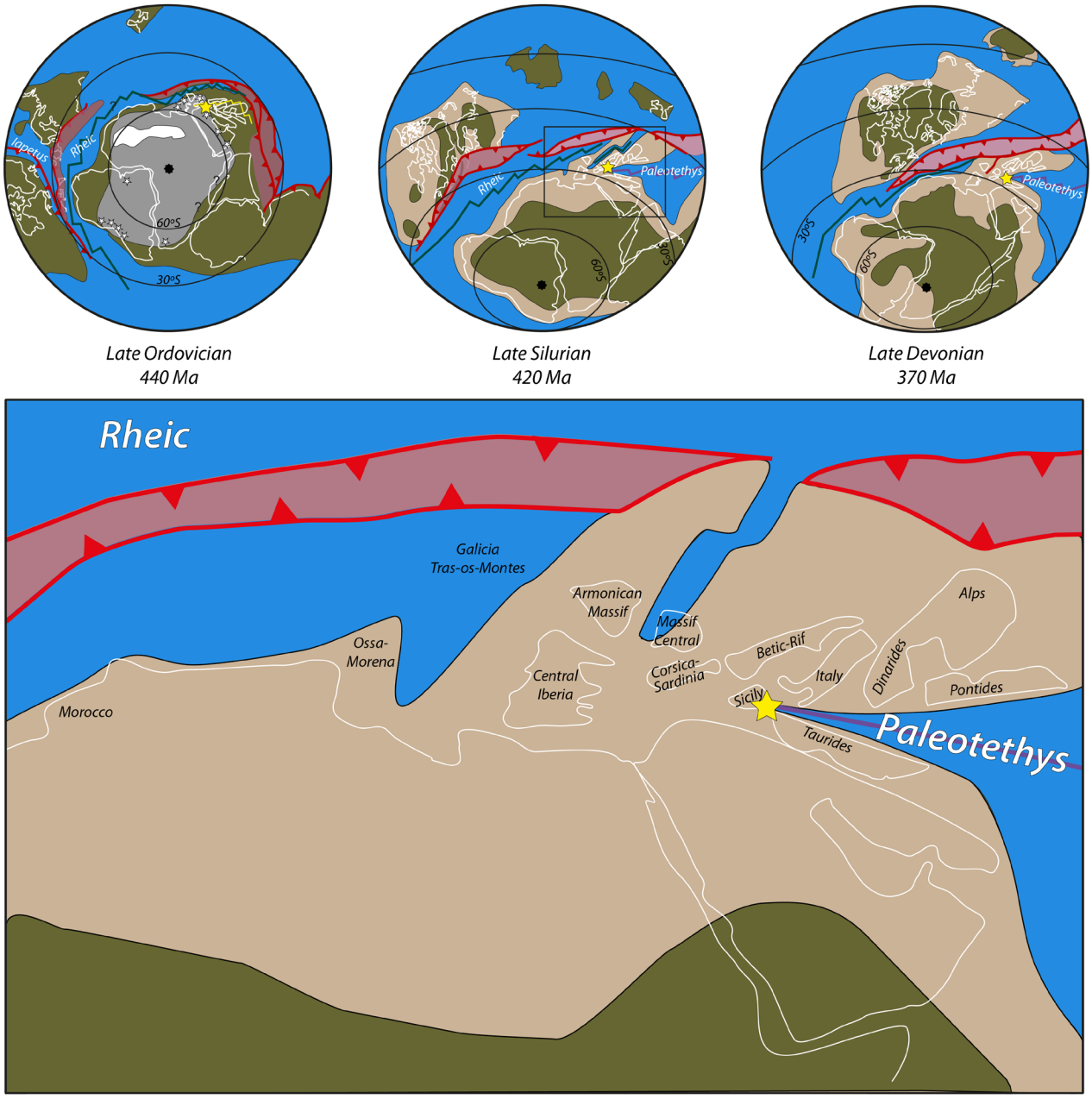


Figure III.5.1. Proposed palaeogeographic sketch map of the Western Mediterranean Intra-Alpine terrains in the southern paleomargin of the European Hunian (Galatian) ribbon continent that was being detached from Northern Gondwana in Silurian times in the westermost embayment of the Paleotethys. The model assumes the position of the Central Iberian terrains in Ordovician time proposed by Bea et al. (2010).

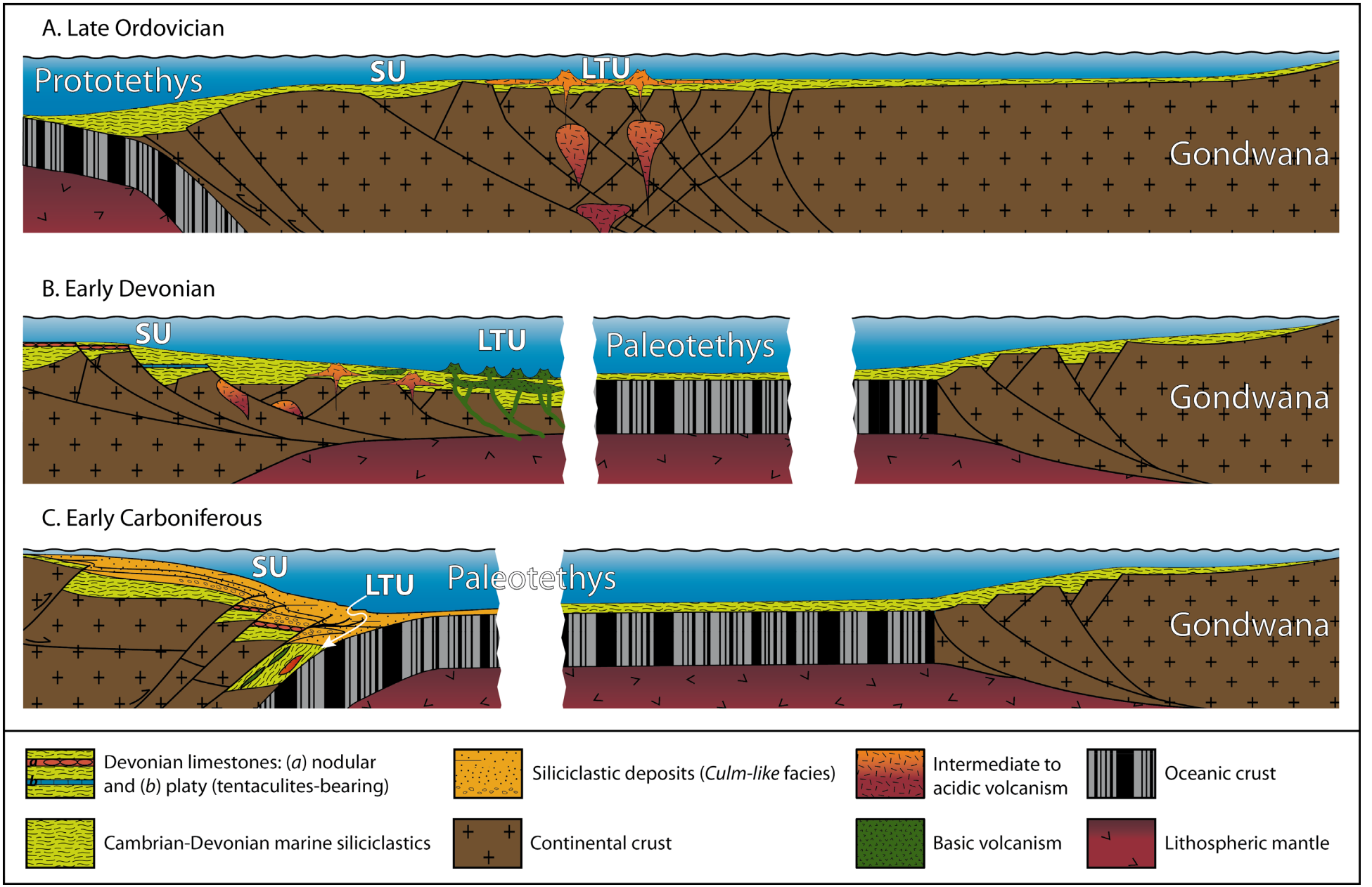


Figure III.5.2. Schematic model for geodynamic and palaeogeographic evolution of the LTU and related terrains in Calabria and other Western Mediterranean Alpine regions. See text for details.

CHAPTER IV
Conclusions

IV. 1. Concluding remarks
IV. 2. Conclusiones

IV.1. Concluding remarks

1) Two groups of Paleozoic successions belonging to the Alpine orogenic belts around the Western Mediterranean have been studied in order to update their stratigraphy and to elucidate their palaeogeographic significance and location in the frame of the Paleotethyan margins of Northern Gondwana. These successions make part of the Malaguide Complex of the Betic Cordillera, and of the Stilo and Longi-Taormina units of the Southern Calabria-Peloritani Terrane. In spite of intense Alpine and pre-Alpine orogenic deformation, these areas still preserve the original stratigraphic features of the Paleozoic successions and, in some intervals, detailed stratigraphic and biostratigraphic studies have been made.

2) As biostratigraphic controls are essential for establishing the chronology and the stratigraphic correlation of the studied successions (mostly because of the monotony of the facies but also due to bad preservation and limited lateral continuity of most of the outcrops) fieldwork was mainly oriented to find limestone horizons suitable for conodont analyses. These analyses, together with lithostratigraphic studies, allowed: i) to identify the main lithostratigraphical features of the studied successions; ii) to date and correlate the studied lithologic units in both the Betic Cordillera and the Calabria-Peloritani Terrane; iii) to identify regionally significant lateral stratigraphic and facies changes; iv) to recognize the palaeogeographical significance of these lateral changes in the frame of the westernmost Paleotethys; and v) to recognize the main features of their geohistorical evolution.

3) Several new stratigraphic horizons have been dated by means of conodonts in the Paleozoic formations of the Malaguide Complex. These are summarized as follows:

3a) Late Ordovician (Hirnantian) conodonts have been found for the first time in the Betic Cordillera. This constitutes the oldest well-dated stratigraphic horizon of the Malaguide Complex up to now, the first Ordovician conodont fauna found in the Western Mediterranean Alpine Orogen, and the youngest Ordovician conodont fauna in Spain.

3b) Early Devonian (Emsian) and Middle Devonian (Eifelian) conodonts have been found for the first time in the Cogollos Vega Zone and the Vélez Rubio-Lorca Corridor, respectively.

3c) Late Devonian conodonts have been found in the Vélez Rubio-Lorca Corridor and, for the first time, in the Cogollos Vega Zone.

3d) Early Carboniferous conodonts have been found in the limestones of the Falcoña Fm. in all the studied areas (Ardales area, Cogollos Vega Zone and Vélez Rubio-Lorca Corridor). The regional extension of the Falcoña Fm. has been confirmed, as well as its importance as correlation horizon, since it has been identified in both all the studied areas and all the Alpine Malaguide units here defined.

3e) Late Carboniferous (Bashkirian) conodonts have been found for the first time in the Betic Cordillera, in the Cortijo de los González area (Epimetamorphic Malaguide Unit). This is also the youngest (hemi) pelagic stratigraphic interval found in the Malaguide Complex. Until now, only shallow-water fauna of Bashkirian age was described in this complex, and it was found in cobbles and boulders of the Marbella Fm.

4) New biostratigraphic data presented in this research for the Paleozoic of the Stilo and the Longi-Taormina units are summarized below:

4a) Early and Middle Devonian conodonts have been described for the first time in the Bivongi-Stilo area of the Stilo Unit.

4b) Frasnian conodonts have been found for the first time in Calabria (San Giovanni 1 and Ponte Vina).

4c) The Upper Kellwasser event has been detected by means of conodonts for the first time

in the Calabria-Peloritani Terrane, and it is there represented by thin anoxic facies (Ponte Vina section).

4d) Several new Famennian successions have been identified and correlated by means of conodonts (San Giovanni 1 and 2, Stilaro and Ponte Vina sections).

4e) Latest Silurian-Early Devonian conodonts have been found for the first time in Sicily (Pizzo Leo sections).

5) The studied outcrops of the Malaguide Complex are all located close to the tectonic contact between the Internal and the External domains. According to both field observations and biostratigraphic information, including those previously published by other authors, the Malaguide Complex has been divided, in the studied areas, into three Alpine tectonic units that also have different Paleozoic successions. These tectonic units have been recognized in different sectors of the Betic Cordillera with similar general features. Therefore, their lateral continuity has been confirmed in all the studied areas, although further studies are needed in order to confirm their presence in the Montes de Málaga area. They are from bottom to top, Epimetamorphic Malaguide Unit, Lower Malaguide Unit, and Upper Malaguide Unit.

6) The Epimetamorphic Malaguide Unit shows a lithostratigraphic succession of undoubtedly Malaguide type. However, the metamorphic overprint that affects it has led to include it within the Alpujarride Complex in other sectors. The only biostratigraphic information obtained from this tectonic Malaguide unit has been found in the Cortijo de los González area. This is a quite particular succession, since faunal associations include Middle Devonian conodonts, only present up to now in the Upper Malaguide Unit, Visean conodonts within the upper member of the Falcoña Fm., and Bashkirian conodonts, found here for the first time in the Malaguide Complex.

7) The González succession is detached from its tectonic substratum, which is constituted by a quite standard Malaguide Paleozoic succession similar to that of the Lower Malaguide Unit. The González succession probably constitutes a tectonic remnant of a previous Malaguide pre-Alpine unit but this hypothesis needs further regional and structural studies in order to be definitely confirmed or rejected.

8) The Alpine Lower Malaguide Unit includes the most typical, thickest and widest outcropping Malaguide succession. It is mainly characterized by the presence of the Calizas alabeadas facies of the Santi Petri Fm. In this Malaguide unit, the only horizons that have been dated by means of conodonts in this study belong to the upper member of the Falcoña Fm. Silurian-Devonian beds were dated by conodonts in the Ardales Castle section, but they essentially made part of a Variscan Upper tectonic Unit. Apart from this site, no Silurian-Devonian conodonts have been found in the Alpine Lower Malaguide Unit.

9) The Paleozoic succession of the Upper Malaguide Unit is the less thick and probably the most complicated in detail from a tectonic point of view: the lateral continuity of their thin Silurian-Devonian stratigraphic horizons of conodont-bearing limestones is frequently interrupted by faults and they mostly appear as wedge-shaped lenses between undifferentiated pelites. Apart from the González outcrop (Epimetamorphic Malaguide Unit) above mentioned, and the Peña del Castillo succession, Silurian-Devonian beds dated by conodonts in this study belong to this Malaguide unit. It is also the only unit that includes the Marbella Fm. in the studied areas.

10) The stratigraphic correlation chart for the Paleozoic of the Malaguide Complex has been confirmed although some clarifications are needed. The typical Malaguide Paleozoic sequence characterizes the lowest units of the Malaguide Complex (Epimetamorphic and, especially, Lower Malaguide Unit) whereas Silurian-Devonian pelagic limestones alternating with pelites and minor sandstone beds replace the Morales and Santi Petri fms. in the Upper Malaguide Unit.

11) Consequently, two types of pre-Carboniferous stratigraphic successions have been identified in the Malaguide Paleozoic. The sequence recorded in the lower units includes turbiditic deposits, such as those of the Morales and Santi Petri fms., with a very probably pre-Devonian and Devonian age, respectively. These facies correspond to distal mixed siliciclastic/carbonatic turbiditic environments deposited within a passive continental margin (slope and basin plain). On the contrary, the sequence recorded in the Upper Malaguide Unit (currently outcropping towards the N in present day coordinates) includes Silurian

and Devonian limestones, in some places nodular, which testify for less deep and less distal (more proximal) environments within the pelagic areas of the same margin.

12) Both sequences record a maximum deepening in the Early Carboniferous, demonstrated by the widespread deposition of the lydites of the Falcoña Fm. and associated Visean limestones.

13) From this time on, the margin became tectonically active. This fact is confirmed by the deposition of siliciclastic turbiditic sediments: the Almogía Fm. (Early Carboniferous), traditionally interpreted as syn-orogenic (Variscan) Culm facies, and the debris flow deposits of the Marbella Fm. (Late Carboniferous).

14) Nevertheless, this syn-orogenic Variscan siliciclastic input was punctuated by carbonate deposition that probably represents short periods of tectonic quiescence during which local deposition of Bashkirian shallow-water and hemipelagic carbonates accounted. The shallow-water carbonates were systematically eroded and have been only recognized as limestone clasts within the Marbella conglomerate Fm. The only outcrop of the coeval hemipelagic Bashkirian limestones is not related, however, to the Marbella Fm. (González outcrop).

15) The current spatial distribution of these Malaguide units is clearly Alpine as they are arranged parallel to the main tectonic contact between the Internal and External domains. Nevertheless, they probably inherited a previous stacking of pre-Alpine tectonic units. Structural analyses performed in the Ardales area on the pre-Alpine structure of the Malaguide Complex indicate a top-to-the-SE directed Variscan deformation.

16) The Paleozoic of the Stilo Unit has been studied in two areas: Stilo-Bivongi (Serre) and Pietrapennata (Aspromonte). The Stilo sequence is very similar to that of the Malaguide Complex and testifies for an equivalent tectono-stratigraphic evolution.

17) The main results were obtained in the Stilo-Bivongi area, where the stratigraphic succession is thickest, more complete and less tectonically disturbed. It includes pre-Devonian pelites and black shales with some limestone intervals found locally, which need further studies as they have not provided yet any identifiable fossil. It also includes Lower, Middle and Upper Devonian beds, Carboniferous lydites and syn-orogenic (Culm-like) siliciclastic deposits.

18) The pre-Devonian deposits are increasingly metamorphic downwards and probably correspond to not very deep but relatively distal marine environments. In the lower part, they include mafic volcanic rocks similar to those of the Longi-Taormina Unit but much less developed. In the upper part, one thin black shale horizon that probably corresponds to Silurian ampelites appear, but it could not be dated.

19) The Devonian of the Stilo-Bivongi area is characterized by an almost continuous pelagic carbonate deposition alternating with pelitic intervals. The presence of several limestone lithofacies during this time points to an evolved continental margin within its pelagic areas, similar to that identified in the Malaguide Complex.

20) Maximum depth of the marine basin recorded in the Stilo Unit took place during Lower Carboniferous as testifies by the presence of lydites above well-dated Famennian limestones. After that, the margin became active and Culm-like facies were deposited.

21) In the outcrops of the Stilo Unit in the Aspromonte area (Staiti-Pietrapennata), only Lower Devonian platy limestones have been dated. Although precise datings could not be obtained by means of conodonts (as only fragments or specimens without biostratigraphic information were found), a lateral change of facies with respect to the Devonian beds of the Stilo-Bivongi area can be inferred. Their facies are slightly different as they are intercalated within thicker pelitic intervals and are dominated by tentaculites limestones, which are also but sporadically present in the Stilo-Bivongi area.

22) The Devonian platy and nodular ("griotte") carbonate lithofacies of the Stilo Unit were deposited in moderately deep environments, whereas the predominantly siliciclastic (turbiditic) lithofacies with abundant basic volcanics and the subordinate carbonate lenses of the Longi-Taormina Unit would have been formed in deeper settings.

23) Paleozoic successions of the Longi-Taormina Unit are mainly characterized by the presence of both calc-alkaline (acidic and intermediate) and alkaline basic volcanism. The

calc-alkaline suite is geochronologically dated as Late Ordovician but basic alkaline volcanism remains yet geochronologically undated. However, basic volcanics outcropping in the Pizzo Leo area have been interpreted as Silurian since they crop out below uppermost Silurian-Lower Devonian limestones first dated by means of conodonts in this study. Paleozoic successions of the Longi-Taormina Unit are characterized also by probably Lower Carboniferous lydites and Culm-like deposits.

24) The deposition of the Paleozoic succession of the Longi-Taormina Unit started under a geodynamic regime related to a Cambrian-Ordovician subduction that produced a Late Ordovician intermediate to acidic volcanism. The generation of this volcanic arc in the Longi-Taormina Unit was coeval to the subduction of the Prototethyan oceanic crust under Northern Gondwana and was associated with a fine-grained siliciclastic sedimentation.

25) As it also accounted in many other sites of the Northern Gondwana margin, persistent Prototethyan subduction along this margin finally determined the extensional collapse of the island arc, and the opening of a back-arc basin, floored by Silurian (and probably Devonian) alkaline mafic volcanics that was related to the opening of a Paleotethyan continental margin in the Longi-Taormina Unit.

26) Such evolution fits well with that known from many central-eastern nord-gondwanic regions during the opening of the Paleotethys. It determined the detachment of a long, ribbon continental crust plate from eastern Northern Gondwana, from which most Paleozoic terrains involved in the Alpine Belt made part.

27) The rifting progressed to ocean spreading during the Silurian and the Devonian in the Alpine areas located towards the E of the Western Mediterranean Alpine belts. Nevertheless, the intracontinental features of the Longi-Taormina Unit mafic volcanism precludes the existence of a true oceanic crust in the Paleozoic terrains of the Western Mediterranean Alpine Belt. This indicates that the Paleotethys ocean closed gradually towards these western regions. The great development of the Longi-Taormina volcanism can be interpreted as a marker of its proximity to true oceanic areas of the Paleotethys, in comparation with the Stilo or Malaguide Paleozoic terrains.

28) So, the Longi-Taormina Unit constituted the most distal area of the Silurian-Devonian continental margin opened to the south of the ribbon continent in the westermost embayment of the Paleotethys, whereas the Stilo Unit represented a more proximal area within the same margin.

29) The Silurian-Devonian Malaguide palaeogeographic zonation reveals an equivalent margin zonation in the westernmost area of the same Paleotethyan margin. In the Malaguide transversal, the proximal area has stratigraphic successions similar to those of the Stilo Unit and belong mainly to the Upper Alpine Malaguide Unit (in addition to the Upper Variscan subunit of the Ardales area and the González succession). The distal zone of the same margin was constituted by the predominantly turbiditic Silurian-Devonian succession of the Lower Malaguide Unit and of the Epimetamorphic Malaguide Unit. Nevertheless the Paleozoic volcanism was very scarce or absent in the distal region of the Malaguide margin, opposited to the palaeogeographic equivalent regions of the Longi-Taormina Unit. This fact reinforces the westward closing of the Paleotethyan oceanic crust.

30) Both the latitudinal drift of whole Gondwanaland and the detachment of the ribbon continent related to the Paleotethys opening determined the northward migration of the Intra-Alpine terrains, from palaeolatitudes related to cold climates and around 50-60ºS during the Hirnantian to lower latitudes in the Silurian-Devonian. This is especially well recorded in the Malaguide successions: the Hirnantian beds are characterized by cool-water conodont associations whereas Silurian, Devonian and especially Carboniferous faunas indicate much warmer climatic conditions.

31) An equivalent gradual change to warmer climates can be inferred for the Calabria-Peloritani Paleotethyan margin. Both in the Stilo and the Longi-Taormina units a sedimentation change from predominantly siliciclastic deposition in this case asociated with volcanoclastic deposition changed gradually to a mixted siliciclastic/carbonatic sedimentation during the Devonian with facies equivalent to those of the Malaguide coeval terrains.

32) The lateral facies evolution in the Silurian-Devonian successions of both the Malaguide Complex and the Calabria-Peloritani Terrane indicates a proximal to distal facies change from NW to SE and N to S, respectively (in present-day coordinates) within the continental margin

succession represented by the studied successions. Consequently, both were located to the N and NW of the Paleotethys. In brief, both Paleozoic Malaguide and Calabria-Peloritani intra-Alpine terrains studied in this research were part of the southern margin of the ribbon continent mentioned before, in the western end of the Paleotethys.

33) In both the Malaguide and the Calabria-Peloritani Paleotethyan margins, the mature (passive) stage of the divergent margin evolution was reached during the earliest Carboniferous, with deposition of radiolarites (lydites) due to the thermal subsidence that affected the whole margin. The opening of the Paleotethys ocean has been completed at the same time but in regions located towards the W, as no traces of ophiolites have been found in the Paleozoic succession of the Western Mediterranean Alpine regions.

34) In mid-Carboniferous times, the northwards Paleotethyan subduction under the Intra-Alpine terrain started. This was the time when the synorogenic Culm-like siliciclastic deposition accounted in both the Stilo and the Longi-Taormina units and in the Malaguide Complex. Consequently, the formerly divergent paleomargin evolved in both areas towards a convergent margin.

35) At the end of the Early Carboniferous and during the Late Carboniferous, the Longi-Taormina Unit and most Paleozoic and older rocks of the Calabria-Pelotirani Terrane, were subjected to Variscan metamorphism. In Late Carboniferous times, the Malaguide terrains were also tectonically deformed but in upper structural levels than the Calabria-Peloritani terrains and immediately after the deposition of the Marbella Fm. in tectonically active deep-marine settings. The stronger metamorphic overprint of the Longi-Taormina Unit terrains, when compared to that recorded in the Stllo and in the Malaguide Complex, was probably related to a shallow underplating of the Longi-Taormina Unit under the piece of the margin represented by the Stilo Unit. This point towards a location of the Calabria-Pelotritani Terrane units closer to the Paleotethyan subduction zone.

36) A widespread Late Carboniferous to Early Permian (300-270 Ma) late-orogenic magmatism generated numerous granitic plutonic bodies in different tectonic units of the CPT (including the Stilo Unit), sealing most of previous Variscan orogenic signatures. Such plutonic stocks have not been found, however, in relation nor to the Paleozoic successions of the Longi-Taormina Unit neither in the Malaguide Complex. This fact also points towards a relatively outer (southern) palaeotectonic location with respect to other tectonic units of the CPT (Stilo Unit in particular). Consequently, the Longi-Taormina Unit was located closer to the Paleotethyan oceanic floor, then in quite distal zones of the formerly divergent passive margin. The absence of Late Carboniferous to Early Permian plutonism in the Malaguide Complex also points to a much more limited subduction in the western studied transversal than in areas located towards the E.

IV.2. Conclusiones

1) Se han estudiado dos grupos de sucesiones paleozoicas pertenecientes al Orógeno Alpino del Mediterráneo Occidental con el fin de actualizar su estratigrafía y de esclarecer su importancia y localización paleogeográfica en el contexto de los márgenes del Paleotethys en el norte de Gondwana. Estas sucesiones forman parte del Complejo Maláguide en la Cordillera Bética y de las Unidades de Stilo y Longi-Taormina en el Terreno Cálabro-Peloritano meridional. A pesar de la intensa deformación Alpina y pre-Alpina, estas áreas todavía conservan las características estratigráficas originales de sus sucesiones paleozoicas y en algunos intervalos se han llevado a cabo estudios estratigráficos y bioestratigráficos de detalle.

2) El trabajo de campo se ha enfocado fundamentalmente hacia la localización de horizontes carbonatados apropiados para el estudio de conodontos, ya que para establecer la cronología y la correlación estratigráfica de las sucesiones estudiadas, estos controles bioestratigráficos son imprescindibles, principalmente debido a la monotonía de las facies, pero también por la mala preservación y la continuidad lateral limitada de la mayoría de los afloramientos. Estos estudios bioestratigráficos, junto con estudios litoestratigráficos, han permitido: i) identificar las características litoestratigráficas principales de las sucesiones estudiadas; ii) datar y correlacionar las unidades litológicas estudiadas tanto en la Cordillera Bética como en el Terreno Cálabro-Peloritano; iii) identificar regionalmente importantes cambios laterales de facies y estratigráficos; iv) reconocer la importancia paleogeográfica de estos cambios laterales de facies en el contexto del Paleotethys occidental; y v) reconocer las características principales de su evolución geohistórica.

3) Se han datado con conodontos nuevos intervalos estratigráficos en las formaciones paleozoicas del Complejo Maláguide:

3a) Se han hallado por primera vez conodontos del Ordovícico Superior en la Cordillera Bética. Esta fauna caracteriza el horizonte más antiguo datado hasta ahora en el Complejo Maláguide, es la primera del Ordovícico hallada en el orógeno alpino del Mediterráneo occidental y la asociación de conodontos ordovícicos más moderna de España.

3b) Se han encontrado por primera vez conodontos del Devónico Inferior (Emsiense) y del Devónico Medio (Eifeliense) en la Zona de Cogollos Vega y el Corredor de Vélez Rubio-Lorca, respectivamente.

3c) Se han encontrado conodontos del Devónico superior en el Corredor de Vélez Rubio-Lorca y por primera vez en la Zona de Cogollos Vega.

3d) Se han encontrado conodontos del Carbonífero Inferior en las calizas de la Formación Falcoña en todas las áreas estudiadas (Ardales, Zona de Cogollos Vega y Corredor de Vélez Rubio-Lorca). Se ha confirmado la extensión regional de la Formación Falcoña, así como su importancia como horizonte de correlación ya que aparece en las distintas áreas estudiadas y en todas las unidades alpinas maláguides que aquí se definen.

3e) Se han encontrado por primera vez conodontos del Carbonífero Superior (Bashkiriense) en la Cordillera Bética, en el área del Cortijo de los González (Unidad Maláguide Epimetamórfica). Este es además el horizonte (hemi) pelágico más joven encontrado en el Complejo Maláguide. Hasta ahora, solo se había descrito fauna marina somera de edad Bashkiriense, en cantos y bloques de la Formación Marbella.

4) Los nuevos datos bioestratigráficos obtenidos en el Paleozoico de la Unidad de Stilo y de la Unidad de Longi-Taormina se resumen a continuación:

4a) Se han encontrado conodontos del Devónico Inferior y Medio por primera vez en la Unidad de Stilo en el área de Stilo-Bivongi (secciones de Assi, y S. Giovanni 1).

4b) Se han hallado por primera vez conodontos de edad Frasniense en Calabria (secciones de S. Giovanni 1 y Ponte Vina).

4c) Se ha detectado el evento Kellwasser con conodontos por primera vez en el Terreno Cálabro-Peloritano, y está representado por finas facies anóxicas (sección de Ponte Vina).

4d) Se han identificado y correlacionado con conodontos varias secciones del Famenniense (secciones de S. Giovanni 1 y 2, Stilaro y Ponte Vina).

4e) Se han encontrado por primera vez conodontos del Silúrico terminal y del Devónico Inferior en Sicilia (área de Pizzo Leo).

5) Todos los afloramientos estudiados en el Complejo Maláguide se encuentran cerca del contacto tectónico entre las Zonas Internas y las Zonas Externas. De acuerdo con las observaciones en el campo y la información bioestratigráfica, incluyendo aquella previamente publicada por otros autores, se ha dividido el Complejo Maláguide, en las áreas estudiadas, en tres unidades tectónicas alpinas que además presentan distintas series paleozoicas. Estas unidades tectónicas se han reconocido en diferentes sectores de la Cordillera Bética con características semejantes. Por lo tanto, se confirma su continuidad lateral aunque su presencia en el área de los Montes de Málaga necesita ser confirmada en futuros estudios. Estas unidades son, de abajo hacia arriba, la Unidad Maláguide Epimetamórfica, la Unidad Maláguide Inferior y la Unidad Maláguide Superior.

6) La Unidad Malaguide Epimetamórfica presenta una sucesión litoestratigráfica claramente de tipo maláguide. Sin embargo, la impronta metamórfica que la afecta ha llevado a incluirla en el Complejo Alpujárride en otros sectores. La única información bioestratigráfica obtenida en esta unidad tectónica maláguide se ha encontrado en el área del Cortijo de los González. Se trata de una sucesión bastante particular, ya que las asociaciones faunísticas incluyen conodontos del Devónico Medio, solo presentes hasta ahora en la Unidad Malaguide Superior, conodontos viseenses en el miembro superior de la Formación Falcoña, y conodontos bashkirienses, que se han encontrado aquí por primera vez en el Complejo Maláguide.

7) La sucesión del Cortijo de los González está despegada de su sustrato original, que está constituido por una sucesión maláguide clásica, equivalente a la de la Unidad Maláguide Inferior. Esta sucesión del Cortijo de los González constituye probablemente un remanente tectónico de una unidad maláguide pre-Alpina, pero esta hipótesis necesita ser confirmada con estudios regionales y estructurales.

8) La unidad alpina inferior incluye la sucesión maláguide más típica, potente y extensa. Se caracteriza principalmente por la presencia de las facies de calizas alabeadas de la Formación Santi Petri. En esta Unidad, los únicos horizontes que se han podido datar con conodontos en este estudio pertenecer al miembro superior de la Formación Falcoña. Los niveles silúrico-devónicos datados con conodontos en el corte de la Peña del Castillo de Ardales forman parte de la Unidad Varisca Superior. Excepto en este lugar, no se han encontrado conodontos silúrico-devónicos en la Unidad Maláguide Alpina Inferior.

9) La serie paleozoica de la Unidad Maláguide Superior es la menos potente y probablemente la más complicada en detalle, desde un punto de vista tectónico: la continuidad lateral de sus delgados horizontes estratigráficos de calizas con conodontos siluro-devónicos se ve interrumpida frecuentemente por fallas y a menudo aparecen como lentejones entre pelitas indiferenciadas. Además de los cortes antes mencionados del Cortijo de los González (Unidad Maláguide Epimetamórfica) y de la Peña del Castillo de Ardales, los niveles siluro-devónicos datados con conodontos en este estudio pertenecen a esta unidad maláguide. Es también la única unidad que incluye la Formación Marbella en las áreas estudiadas.

10) El cuadro de correlación estratigráfica del Paleozoico Maláguide se ha confirmado aunque son necesarias algunas puntualizaciones. Las unidades inferiores del Complejo Maláguide (Unidad Epimetamófica y especialmente, la Unidad Maláguide Inferior) están caracterizadas por la típica serie del paleozoico maláguide mientras que las calizas pelágicas siluro-devónicas con alternancias pelíticas y algunos niveles de areniscas ocupan la posición de las formaciones Morales y Santi Petri en la Unidad Malaguide Superior.

11) Por lo tanto, se han identificado dos tipos de sucesiones pre-carboníferas en el paleozoico maláguide. La secuencia registrada en las unidades inferiores incluye depósitos turbidíticos como los de las formaciones Morales y Santi Petri, con una

edad probablemente pre-devónica y devónica, respectivamente. Estas facies corresponden con ambientes distales siliciclástico-carbonatados depositados en un margen continental pasivo (talud y cuenca). Por el contrario, la secuencia registrada en la Unidad Maláguide Superior (que aflora actualmente al norte en coordenadas actuales) incluye calizas silúricas y devónicas, en algunos lugares nodulosas, que apuntan a ambientes menos profundos y menos distales dentro de las áreas pelágicas del mismo margen.

12) Ambas secuencias registran una máxima profundización en el Carbonífero Inferior, atestiguada por el depósito generalizado de liditas de la Formación Falcoña y calizas viseenses asociadas.

13) A partir de este momento el margen se volvió activo. Este hecho queda confirmado por el depósito de los sedimentos turbidíticos siliciclásticos de la Formación Almogía (Carbonífero Inferior), interpretadas tradicionalmente como facies sinorogénicas variscas de tipo Culm, y los depósitos de flujos de derrubios de la Formación Marbella (Carbonífero Superior).

14) No obstante, este aporte siliciclástico sinorogénico se vio interrumpido por un depósito carbonatado que probablemente aconteció durante cortos periodos de inactividad tectónica durante los cuales tuvo lugar el depósito de carbonatos bashkirienses de aguas someras y hemipelágicos. Los carbonatos de aguas someras fueron erosionados en su totalidad y solo se han reconocido como clastos dentro de conglomerados de la Formación Marbella. Sin embargo, el único afloramiento de calizas hemipelágicas bashkirienses contemporáneas no está relacionado con la Formación Marbella (corte de los González).

15) La distribución espacial actual de estas unidades maláguides es claramente alpina ya que están distribuidas paralelamente al contacto entre las Zonas Internas y las Zonas Externas. Sin embargo, la estructura del Complejo Maláguide probablemente ha heredado un apilamiento previo de unidades tectónicas pre-alpinas. Los análisis estructurales de la estructura pre-alpina del Complejo Maláguide llevados a cabo en el área de Ardales indican una deformación varisca dirigida hacia el SE.

16) El Paleozoico de la Unidad de Stilo se ha estudiado en dos áreas: Stilo-Bivongi (Macizo de la Serre) y región de Staiti-Pietrapennata (Macizo del Aspromonte). La secuencia de Stilo es similar a la del Complejo Maláguide e indica una evolución tectono-estratigráfica equivalente.

17) Los resultados más importantes se han obtenido en el área de Stilo-Bivongi, donde la sucesión estratigráfica es más potente, más completa y está menos deformada. Esta secuencia incluye pelitas pre-devónicas y pizarras negras con algunos horizontes calizos que necesitan ser estudiados más en profundidad ya que no han liberado ningún fósil identificable. También incluye calizas del Devónico Inferior, Medio y Superior, liditas carboníferas y depósitos sinorogénicos siliciclásticos (tipo Culm).

18) Los depósitos pre-devónicos son gradualmente más metamórficos hacia abajo y probablemente se depositaron en ambientes marinos no muy profundos pero relativamente distales. En su parte inferior incluyen rocas volcánicas máficas equivalentes a las de la Unidad de Longi-Taormina, pero mucho menos desarrolladas. En la parte superior, estos depósitos pre-devónicos incluyen un fino horizonte de pizarras negras que probablemente corresponden con ampelitas silúricas, pero no ha podido datarse.

19) El Devónico del área de Stilo-Bivongi se caracteriza por una sedimentación pelágica carbonatada casi continua con intervalos pelíticos. La presencia de varias litofacies calcáreas durante este periodo apunta hacia un margen continental evolucionado en sus áreas pelágicas, equivalente al identificado en el Complejo Maláguide.

20) La profundización máxima, indicada por la presencia de liditas sobre calizas famennienses bien datadas, de la cuenca registrada en la Unidad de Stilo tuvo lugar durante el Carbonífero Inferior.

21) En los afloramientos de la Unidad de Stilo en el área del Aspromonte (Staiti-Pietrapennata), solo se han podido datas calizas en plaquetas del Devónico Inferior. Aunque no se han podido obtener dataciones precisas mediante conodontos (ya que solo se han encontrado fragmentos o ejemplares sin información bioestratigráfica), puede deducirse un cambio lateral de facies con respecto a las rocas devónicas del área de Stilo-Bivongi. Sus facies son ligeramente distintas, ya que están intercaladas entre intervalos pelíticos más gruesos y están dominados por calizas de tentaculites, que solo aparecen esporádicamente

en el área de Stilo-Bivongi.

22) Las litofacies carbonatadas en plaquetas y nodulosas ("griotte") del Devónico de la Unidad de Stilo se depositaron en ambientes moderadamente profundos, mientras que la litofacies predominantemente siliciclástica (turbidítica) con abundantes vulcanitas básicas y algunas lentes carbonatadas de la Unidad de Longi-Taormina se habrían depositado en ambientes más profundos.

23) Las sucesiones paleozoicas de la Unidad de Longi-Taormina se caracterizan principalmente por la presencia de vulcanismo tanto calco-alcalino (ácido e intermedio) como alcalino básico. El edificio calco-alcalino está datado geocronológicamente como Ordovícico Superior pero el vulcanismo básico alcalino no se ha datado geocronológicamente todavía. Sin embargo, las vulcanitas básicas que afloran en el área de Pizzo Leo han sido interpretadas como silúricas ya que aparecen por debajo de calizas del Silúrico-Devónico Inferior datadas con conodontos por primera vez en este estudio. Las sucesiones paleozoicas de la Unidad de la Longi-Taormina se caracterizan también por liditas de probable edad Carbonífero Inferior y depósitos tipo Culm.

24) El depósito de las sucesiones paleozoicas de la Unidad de Longi-Taormina comenzó en un régimen geodinámico relacionado con la subducción cambro-ordovícica que dio lugar a un vulcanismo de intermedio a ácido de edad Ordovícico Superior. La generación de este arco volcánico en la Unidad de Longi-Taormina fue contemporánea a la subducción de la corteza oceánica del Prototethys bajo Gondwana septentrional, y estuvo asociada a una sedimentación siliciclástica de grano fino.

25) La continua subducción prototethysiana a lo largo de este margen, como también ocurrió en otras áreas del margen nord-gondwánico, determinó finalmente el colapso extensional del arco isla y la apertura de una cuenca de retro-arco inundada por rocas volcánicas máficas alcalinas silúricas (y probablemente devónicas) relacionadas con la apertura de un margen continental paleotethysiano en la Unidad de Longi-Taormina.

26) Esta evolución encaja bien con la que se conoce para otras muchas regiones nord-gondwánicas durante la apertura del Paleotethys. Esto determinó la separación de una placa de corteza continental estrecha y alargada ("costilla continental" o ribbon continent) desde el norte de Gondwana, de la cual derivan la mayoría de los terrenos paleozoicos del cinturón alpino.

27) El rifting progresó con la expansión del fondo oceánico durante el Silúrico y el Devónico en las áreas alpinas localizadas al Este del Orógeno Alpino del Mediterráneo Occidental. Sin embargo, las características intracontinentales del vulcanismo máfico de la Unidad de Longi-Taormina excluye la existencia de verdadera corteza oceánica en los terrenos paleozoicos estudiados. Esto indica que el océano Paleotethys se cerraba gradualmente hacia las regiones occidentales. El gran desarrollo de vulcanismo en la Unidad de Longi-Taormina puede interpretarse, todo lo más, como un indicador de su proximidad a verdaderas áreas oceánicas del Paleotethys, en comparación con la Unidad de Stilo o los terrenos del paleozoico maláguide.

28) Por lo tanto, la Unidad de Longi-Taormina constituye el área más distal del margen continental que se abrió al sur de la costilla continental separada de Gondwana, mencionada más arriba, en la terminación occidental del Paleotethys, mientras que la Unidad de Stilo representa un área más proximal dentro del mismo margen.

29) Una zonación paleogeográfica semejante se observa también, durante el Silúrico y el Devónico, en la transversal más occidental dentro del mismo margen del Paleotethys. Así, en el transecto maláguide, el área proximal está formada por sucesiones estratigráficas equivalentes a las de la Unidad de Stilo y pertenecen principalmente a la Unidad Malaguide Superior (además de la subunidad varisca superior en el área de Ardales y la sucesión de los González). La zona más distal del mismo margen estuvo constituida por las sucesiones siluro-devónicas predominantemente turbidíticas de la Unidad Maláguide Inferior y de la Unidad Maláguide Epimetamórfica. Sin embargo, el vulcanismo paleozoico fue muy escaso o ausente en la región distal del margen maláguide, en oposición a las regiones paleogeográficas equivalentes de la Unidad de Longi-Taormina. Este hecho refuerza la hipótesis arriba enunciada acerca de la terminación occidental de la corteza oceánica del Paleotethys.

30) La deriva latitudinal de toda Gondwana y el despegue de la costilla continental relacionados con la apertura del Paleotethys determinó la migración de los terrenos intra-Alpinos desde altas paleolatitudes meridionales relacionadas con climas fríos, y en torno a 50-60ºS durante el Ordovícico, a latitudes más bajas en el Silúrico y el

Devónico. Esto está especialmente bien registrado en las sucesiones maláguides: el Hirnantiense está caracterizado por asociaciones de conodontos de aguas frías mientras que las faunas del Silúrico, del Devónico y especialmente del Carbonífero indican condiciones climáticas mucho más cálidas.

31) Un cambio gradual equivalente hacia climas más cálidos puede deducirse también para el paleomargen cálabro-peloritano. Tanto en la Unidad de Stilo como en la Unidad de Longi-Taormina, la sedimentación de depósitos predominantemente siliciclásticos, en este caso asociados con depósitos vulcanoclásticos, cambió gradualmente a una sedimentación mixta siliciclástico-carbonatada durante el Devónico con facies equivalentes a las de los terrenos maláguides contemporáneos.

32) La evolución lateral de facies en las sucesiones siluro-devónicas del Complejo Maláguide y del Terreno Cálabro-Peloritano desvela una zonación proximal-distal desde el NO hacia el SE y del norte al sur, respectivamente (en coordenadas actuales) dentro del margen continental representado en las dos transversales estudiadas. Por lo tanto, ambas estuvieron localizadas al N y al NO del Paleotethys. En definitiva, los terrenos paleozoicos intra-Alpinos maláguide y cálabro-peloritanos estudiados en este trabajo formarían parte del margen meridional de la costilla continental desgajada de Gondwana antes mencionada, en la terminación occidental del Paleotethys.

33) Tanto en el margen maláguide como en el cálabro-peloritano, la etapa de madurez (pasiva) de la evolución de margen continental divergente se alcanzó durante el Carbonífero inicial con el depósito de radiolaritas (liditas) debido a la subsidencia termal que afectó a todo el margen. La apertura del océano Paleotethys se completó al mismo tiempo, pero en regiones localizadas hacia el Este, ya que no se han encontrado restos de ofiolitas en las sucesiones paleozoicas de los dominios alpinos del Mediterráneo occidental.

34) Durante el Carbonífero medio comenzó la subducción hacia el norte del Paleotethys bajo los terrenos intra-Alpinos. En este momento comenzó el depósito sinorogénico tipo Culm, tanto en la Unidad de Stilo y en la Unidad de Longi-Taormina como en el Complejo Maláguide. Por lo tanto, el antiguo paleomargen divergente evolucionó en ambas áreas hacia un margen convergente.

35) Al final del Carbonífero Inferior y durante el Carbonífero Superior, la Unidad de Longi-Taormina y la mayoría de las rocas paleozoicas y más antiguas del Terreno Cálabro Peloritano fueron sometidas a metamorfismo varisco. Durante el Carbonífero Superior inmediatamente después del depósito de la Formación Marbella en ambientes marinos profundos orogénicamente activos, los terrenos maláguides fueron también deformados tectónicamente pero en niveles estructurales superiores a los de los terrenos cálabro-peloritanos. La impronta metamórfica más fuerte en los terrenos de la Unidad de Longi-Taormina, en comparación con la registrada en la Unidad de Stilo o en el Complejo Maláguide, estuvo probablemente relacionada con la subcreción tectónica (underplating) a poca profundidad de esta unidad bajo el fragmento del margen de la costilla continental representado por la Unidad de Stilo. Con respecto al Complejo Maláguide, esto apunta hacia una localización de las unidades del Terreno Cálabro-Peloritano más cercana a la zona de subducción del Paleotethys.

36) El magmatismo tardi-orogénico generalizado durante el Carbonífero Superior y el Pérmico Inferior (300-270 Ma) generó numerosos cuerpos plutónicos en diferentes unidades tectónicas del Terreno Cálabro-Peloritano (incluyendo la Unidad de Stilo), que sellaron la mayoría de las señales orogénicas previas. Sin embargo, estos cuerpos plutónicos no se han encontrado en relación con las sucesiones paleozoicas del Complejo Maláguide ni de la Unidad de Longi-Taormina. Este hecho también apunta hacia una posición paleotectónica relativamente más externa (meridional) de la Unidad de Longi-Taormina con respecto a otras unidades tectónicas del Terreno Cálabro-Peloritano (Unidad de Stilo en particular). Por lo tanto, la Unidad de Longi-Taormina estuvo localizada más cerca del fondo oceánico del Paleotethys, en zonas bastante distales del antiguo paleomargen pasivo divergente. La ausencia de plutonismo del Carbonífero Superior al Pérmico Inferior en el Complejo Maláguide también apunta hacia una subducción mucho más limitada en la transversal occidental estudiada que en áreas localizadas hacia el este.

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APPENDIX I

Main publications related to this Ph.D. Thesis

Navas-Parejo, P., Rodríguez-Cañero, R. and Martín-Algarra, A. (2008). Conodontos del Carbonífero maláguide en Vélez Rubio (Cordillera Bética, SE España). *Geotemas*, 10. 1273-1276.

Navas-Parejo, P., Somma, R., Martín-Algarra, A., Perrone, V. and Rodríguez-Cañero, R. (2009a). First record of Devonian orthoceratid-bearing limestones in southern Calabria (Italy). *Comptes Rendus Palevol*, 8. 365-373.

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Martín-Algarra, A., Mazzoli, S., Perrone, V., Rodríguez-Cañero, R. and **Navas-Parejo, P.** (2009). Variscan Tectonics in the Malaguide Complex (Betic Cordillera, Southern Spain): Stratigraphic and Structural Alpine versus Pre-Alpine Constraints from the Ardales Area (Province of Malaga). I. Stratigraphy. *Journal of Geology*, 117. 241–262.

Rodríguez-Cañero, R., Martín-Algarra, A., Sarmiento, G.N. and **Navas-Parejo, P.** (2010). First Late Ordovician conodont fauna in the Betic Cordillera (South Spain): a palaeobiogeographical contribution. *Terra Nova*, 22. 330–340.

Navas-Parejo, P., Martín-Algarra, A. and Martínez-Pérez, C. (2011). Primeros datos sobre la presencia de conodontos del Emsiense (Devónico Inferior) en el Complejo Maláguide de la provincia de Granada. *Paleontologia i evolució, Memòria especial*, 5. 425-429.

Navas-Parejo, P., Martín-Algarra, A. and Rodríguez-Cañero, R. (2012a). Conodontos del Carbonífero maláguide de la región de Ardales (Cordillera Bética occidental). In: *Paleodiversity and Paleoecology of Iberian Ecosystems* (Martínez-Pérez *et al*. (eds.), 52-54 (Sot de Chera, Valencia).

Navas-Parejo, P., Rodríguez-Cañero, R. and Martín-Algarra, A. (2012b). Primer registro de un horizonte estratigráfico hemipelágico con conodontos del Carbonífero Superior en el Complejo Maláguide oriental (Cordillera Bética). *Geogaceta*, 52.

Conodontos del Carbonífero maláguide en Vélez Rubio (Cordillera Bética, SE Spain)

Malaguide Carboniferous Conodonts from Velez Rubio (Betic Cordillera, SE Spain)

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Resumen: La presencia de la Formación Falcoña (Carbonífero Inferior: Mississippiense) ha podido ser demostrada por debajo de materiales turbidíticos de facies Culm correlacionables con la Formación Almogía en varios afloramientos del Complejo Maláguide del área de Vélez Rubio. En dos de las secciones estudiadas, la Formación Falcoña está constituida por un miembro inferior de radiolaritas negras (liditas) y un miembro superior de calizas pelágicas grises-azuladas, al igual que se observa en el corte tipo de la formación en el sector occidental del Complejo Maláguide. Las calizas han proporcionado conodontos del Viseense superior, los primeros de esta edad determinados en el Complejo Maláguide del sector oriental. Los nuevos datos lito y bioestratigráficos obtenidos confirman que la Formación Falcoña se depositó durante un periodo de sedimentación pelágica generalizada, profunda y muy uniforme en todos los sectores del Dominio Maláguide. Esto sucedió tras el rifting siluro-devónico que afectó a las regiones nororientales de Gondwana y que culminó con la apertura del Paleotethys. A continuación se inició la orogenia Varisca que culminó durante el Carbonífero Superior (Pennsylvaniense) con el cierre de las cuencas del extremo occidental del Paleotethys y la formación de la Pangea.

Palabras clave: conodontos, bioestratigrafía, Mississippiense, Complejo Maláguide, Formación Falcoña.

Abstract: *The Falcoña Formation (Lower Carboniferous: Mississippian), defined in the western sector of the Malaguide Complex, has been found in several outcrops of the eastern Malaguide Complex (Vélez Rubio area) below turbiditic successions correlatable to the Almogía Formation. As well as in the type section, two sections of this formation studied near Vélez Rubio are made of two members: black radiolarites (lydites) below, and gray-bluish pelagic limestones above. These limestones provided the first conodonts of Late Viséan age found in the eastern outcrops of the Malaguide Complex. The new litho- and biostratigraphic data confirm that, also in these eastern regions of the Malaguide Domain, the Falcoña Formation was deposited during a period of generalized deep and very uniform pelagic sedimentation. This accounted after the silurian-devonian rifting of the northeastern Gondwana regions, which culminated with the opening of the deep marine paleotethyan basins. After deposition of the Falcoña Formation, the Variscan orogeny started with sedimentation of thick turbiditic succesions of the Almogía Formation (Culm facies), and culminated during the Late Carboniferous (Pennsylvanian), at the same time that other deep basins of the Westernmost Paleotethys were closed during Pangea amalgamation.*

Key words: *conodonts, biostratigraphy, Mississippian, Malaguide Complex, Falcoña Formation.*

INTRODUCCIÓN

El Paleozoico de la Cordillera Bética aflora casi exclusivamente en las Zonas Internas y suele estar formado por rocas metamórficas en la mayoría de los casos. Por ello, las dataciones en estos terrenos son muy escasas y, en su mayor parte, se han obtenido en los afloramientos occidentales del Complejo Maláguide. El Maláguide es el complejo tectónico superior de la pila de mantos de corrimiento que conforma las Zonas Internas de la Cordillera Bética.

En el sector oriental, el Paleozoico del Complejo Maláguide aparece ampliamente representado en el Corredor de Vélez Rubio (Fig. 1). En el área de Chirivel, Van den Boogaard (1965) dató un Silúrico superior y un Devónico superior y proporcionó una relación de las especies encontradas e ilustraciones de dos de ellas. Geel (1973) detectó la presencia de un Devónico inferior por el hallazgo de un espécimen de un posible *Spathognathodus frankenwaldensis* y proporciona el dato de la existencia de un Viseense inferior alto - Viseense superior, datado con conodontos, pero no detalla las especies encontradas.

El objetivo de este trabajo es dar a conocer nuevos resultados sobre la estratigrafía del Paleozoico del Complejo Maláguide en el área de Vélez Rubio y, en particular, presentar los hallazgos de conodontos del Carbonífero obtenidos en esta región.

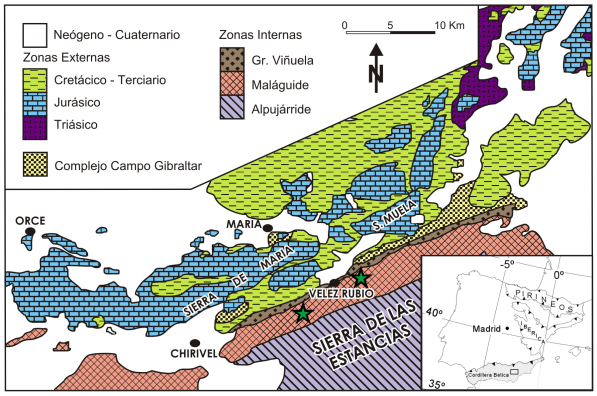

FIGURA 1. *Esquema tectónico del Complejo Maláguide en el Corredor de Vélez Rubio.*

ESTRATIGRAFÍA

El Paleozoico maláguide de Vélez Rubio fue incluido por Soediono (1971) en la formación Piar, un conjunto pizarroso-grauváquico de origen turbidítico, con niveles conglomeráticos, con delgadas intercalaciones pelágicas silíceas y calcáreas y atravesado en algunas zonas por diques de rocas subvolcánicas. Esta formación fue elevada al rango de grupo (Grupo Piar) (Martín-Algarra, 1987, *in*: Martín-Algarra y Rodríguez-Cañero, 2004) al compararla con los terrenos del Complejo Maláguide del sector occidental. Allí se diferencian cinco formaciones: Morales, Santi Petri, Falcoña, Almogía y Marbella, cuya estratigrafía y nomenclatura ha sido recientemente actualizada por Martín-Algarra y Rodríguez-Cañero (2004).

En el corte tipo de la Formación Falcoña se reconocen un miembro inferior de liditas, que ha proporcionado radiolarios del Tournaisiense, y un miembro superior de calizas pelágicas con conodontos del Viseense (Rodríguez-Cañero y Guerra-Merchán, 1996; O'Dogherty *et al*., 2000, *in*: Martín-Algarra y Rodríguez-Cañero, 2004).

La Formación Falcoña está presente en muchos cortes estudiados en la región de Vélez Rubio. Sin embargo, en la mayoría de los casos únicamente aparece el miembro inferior de liditas. Sólo en las dos secciones estudiadas en este trabajo se han encontrado las calizas del miembro superior (Fig. 2). La primera sección aflora a lo largo de un tributario del barranco de Fuente Alegre, al ENE de Vélez Rubio (N de la prov. de Almería). La segunda se localiza en el barranco Pedrero y en ella Geel (1973) señaló la existencia de conodontos del Viseense, sin mencionar la asociación. Las calizas muestreadas en el barranco de Fuente Alegre afloran formando un anticlinal de charnela aguda, con dirección axial NE-SO y vergencia al SE y en el barranco Pedrero lo hacen en forma de lentejón calcáreo con estratificación vertical que probablemente corresponde a una charnela sinclinal isoclinal.

La posición estratigráfica de los niveles calizos muestreados es la misma en ambos casos: a techo de las liditas y a muro de una formación tipo Culm (Formación Almogía). En las dos secciones estudiadas la sucesión estratigráfica es comparable; de muro a techo aparecen (Fig. 2):

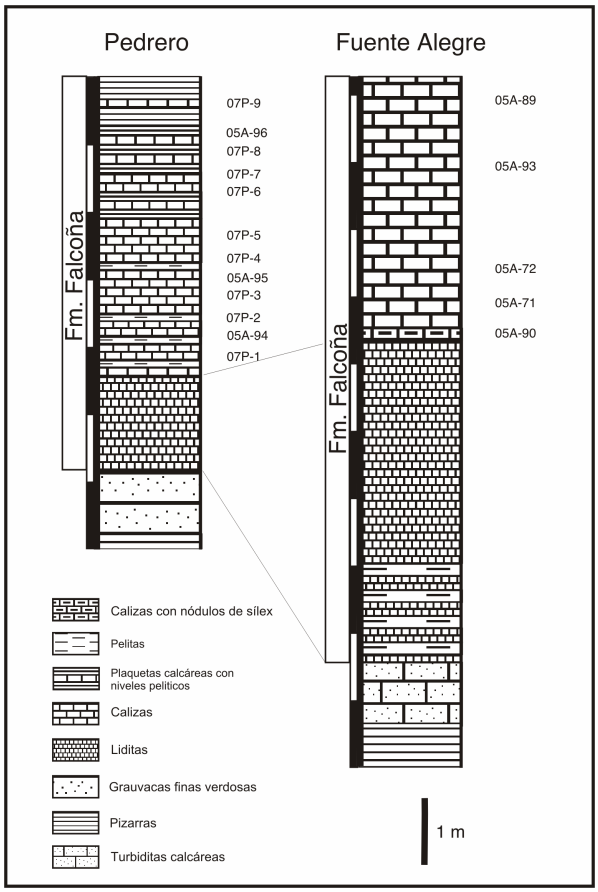

FIGURA 2. *Columnas estratigráficas sintéticas de los afloramientos estudiados.*

1) Pizarras y areniscas turbidíticas de grano fino. En Fuente Alegre aparecen también niveles carbonatados, que son más abundantes hacia la parte alta, donde también aparecen calcarenitas y microconglomerados con abundantes terrígenos (cuarzo, sobre todo).

2) Radiolaritas finamente estratificadas, fuertemente replegadas y recristalizadas, de colores oscuros (liditas). Localmente, en Fuente Alegre, intercalan pizarras y grauvacas de grano fino a muy fino.

3) Calizas grises, en plaquetas, micríticas y localmente con nódulos de sílex, también fuertemente replegadas y recristalizadas, cuyo espesor original apenas alcanza tres metros. En Pedrero presentan intercalaciones pelíticas más abundantes hacia techo. Las muestras para conodontos se han cogido en este tramo. Concretamente se han recogido once muestras en el afloramiento de Fuente Alegre y dieciocho en el de Pedrero.

4) Pizarras y grauvacas con estructuras turbidíticas, de facies Culm. En Fuente Alegre también aparecen abundantes niveles de conglomerados.

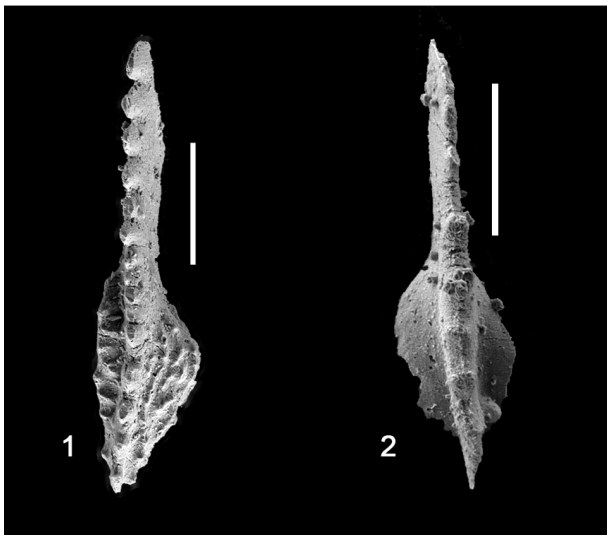

FIGURA 3. *Las escalas gráficas equivalen a 200µm.* ***1.*** *Gnathodus bilineatus Roundy. Elemento Pa 05A-89-12. Fuente Alegre.* ***2.*** *Lochriea commutata Branson y Mehl. Elemento Pa 05A-71-3. Fuente Alegre.*

BIOESTRATIGRAFÍA

De las once muestras recogidas en Fuente Alegre, sólo cinco han proporcionado elementos conodontales reconocibles. En el caso de Pedrero, se recogieron dieciocho muestras, siendo productivas doce de ellas. Todos los ejemplares hallados se encuentran mal conservados. La mayor parte de los ejemplares encontrados y determinados corresponden a elementos Pa de *Gnathodus bilineatus* (Roundy, 1926), que aparece en el Viseense superior y llega hasta el Serpukhoviense inferior (Perret, 1993). En cuatro de las muestras de Fuente Alegre y en tres de Pedrero, además, coexiste con *Lochriea commutata* (Branson y Mehl, 1941), lo que indica una edad Viseense superior, y más concretamente zona de bilineatus – commutatus, (*sensu* Perret, 1993).

SISTEMÁTICA

Gnathodus bilineatus (Roundy, 1926)

Fig. 3, 1.

*1926 *Polygnathus bilineatus* Roundy, lám. 3, figs. 10a-c.
1993 *Gnathodus bilineatus* Roundy; Perret, 329, lám. 110, figs. 11, 12, 15-31.
2005 *Gnathodus bilineatus* Roundy; Blanco-Ferrera *et al*., 22, fig. 6, 17-21.

Los detalles morfológicos que se utilizaron para caracterizar el elemento Pa de esta especie fueron: la posesión de un parapeto interno largo, un gran desarrollo de los nódulos sobre la cúpula externa y una carena ensanchada y con una fila de nódulos adyacentes en su parte posterior. No obstante, diversos autores han puesto de manifiesto la gran variabilidad que presentan estos caracteres, dependiendo de la fase evolutiva, de la etapa de crecimiento y, quizás también, de factores ecológicos.

Lochriea commutata (Branson y Mehl, 1941)

Fig. 3, 2.

*1941 *Spathognathodus commutatus* Branson y Mehl, 98, lám. 19, figs. 1-4.
1985 *Paragnathodus commutatus* Branson y Mehl; Belka, 46, lám. 11, figs. 7-8.
2005 *Lochriea commutata* Branson y Mehl; Blanco-Ferrera *et al*., 22, fig. 6, 2-3.

La cúpula de este elemento carece de ornamentación y, en vista superior, tiene un contorno de ovalado a subcuadrado. Los dentículos de la carena tienden a ensancharse lateralmente, aunque no tanto como en el caso de *Paragnathodus cracoviensis* Belka, del cual se diferencia por presentar este último una lámina con un ensanchamiento lateral muy fuerte, sobre la que se desarrollan puentes transversales.

DISCUSIÓN Y CONCLUSIONES

Los datos obtenidos demuestran la presencia del Viseense superior en el Carbonífero maláguide del área de Vélez Rubio y, por tanto, confirman la pertenencia de las calizas y las liditas suprayacentes de los cortes estudiados a la Fm. Falcoña (O'Dogherty *et al*., 2000, *in* Martín-Algarra y Rodríguez-Cañero, 2004). También ayudan a precisar las correlaciones entre el sector oriental y el occidental del Paleozoico del Complejo Maláguide.

En ambas secciones, las pizarras y areniscas turbidíticas infrayacentes a la Fm. Falcoña son equivalentes a la Fm. Santi Petri. Las pizarras, grauvacas y areniscas de facies Culm corresponden a la Fm. Almogía y los conglomerados asociados son semejantes a los del Miembro Retamares del área tipo.

El depósito de la Formación Falcoña coincidió con un periodo de sedimentación pelágica generalizada, profunda y muy uniforme en todos los sectores del Dominio Maláguide. Tras el rifting siluro-devónico que afectó a las regiones nororientales de Gondwana y que culminó con la apertura del Paleotethys, se produjo una profundización y homogeneización de las facies al generalizarse la subsidencia térmica durante el Mississippiense.

Coincidiendo con una situación de nivel del mar alto (*highstand*), la Formación Falcoña se depositó en una cuenca subalimentada de terrígenos. La cuenca se localizaba en el extremo occidental del Paleotethys a una latitud ecuatorial y tenía una alta productividad planctónica. Durante el Tournaisiense el miembro radiolarítico se depositó en un ambiente de sedimentación pelágico profundo situado por debajo del nivel de compensación de la calcita (CCD). Durante el Viseense se produjo un descenso del CCD y la sedimentación pasó a ser calcárea.

El depósito de la Formación Falcoña precedió al cierre de las cuencas del Paleotethys durante la orogenia Varisca en el Dominio Maláguide. Poco después, a partir del inicio del Serpukhoviense, comenzó la sedimentación sinorogénica de facies Culm (Formación Almogía).

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General palaeontology (Biostratigraphy)

First record of Devonian orthoceratid-bearing limestones in southern Calabria (Italy)

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Abstract

Lower to Middle Devonian orthoceratid-bearing nodular limestones in Calabria are described here for the first time, along the Fiumara Assi section. The succession is tectonically inverted and has been dated by conodonts. The lower beds are Lochkovian–Lower Pragian, because they provided *Icriodus* cf. *steinachensis* and *Pelekysgnathus serratus*, which occur in the delta-sulcatus Zones. Upwards appear *Polygnathus* cf. *dehiscens*, corresponding to the dehiscens-gronbergi Zones, and *Ozarkodina* cf. *steinhornensis miae* of the dehiscens-inversus Zones, both indicating a topmost Pragian–Lower Emsian age. The Eifelian (or younger) age for the top of the succession is demonstrated by *Polygnathus linguiformis linguiformis*. This succession, as well as similar coeval deposits of the southern Variscan Chain (southwestern Sardinia, eastern Pyrenees), Betic–Rifian Maláguides-Ghomarides and southern Alps, made part of a western embayment of the Palaeotethys. ***To cite this article: P. Navas-Parejo et al., C. R. Palevol 8 (2009).***
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Résumé

Présence de calcaires à orthoceratidés du Dévonien en Calabre méridionale (Italie). Le Dévonien inférieur et moyen à faciès de calcaires noduleux à orthoceratidés a été trouvé pour la première fois en Calabre, dans la coupe de la Fiumara Assi. La succession est renversée (tectonique varisque) et a été datée par des conodontes. La base de la succession est Lochkovien–Pragien inférieur, datée par *Icriodus* cf. *steinachensis* et *Pelekysgnathus* gr. *serratus*, appartenant aux zones delta-sulcatus. Les niveaux successifs, à *Polygnathus* cf. *dehiscens*, correspondant aux zones dehiscens-gronbergi et *Ozarkodina* cf. *steinhornensis miae* des zones dehiscens-inversus, indiquent un âge Pragien–Emsien. L'âge Eifelien (ou éventuellement plus jeune) de la partie supérieure de la coupe est démontrée par *Polygnathus linguiformis linguiformis*. Cette succession, comporte des dépôts comparables à ceux de la Chaîne varisque méridionale (Sardaigne sudoccidentale, Pyrénées orientales), des Malaguides-Ghomarides béticorifaines et des Alpes méridionales et faisait partie de la terminaison occidentale de la Paléotethys. ***Pour citer cet article : P. Navas-Parejo et al., C. R. Palevol 8 (2009).***
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Keywords: Stratigraphy; Conodonts; Palaeozoic; Stilo Unit; Calabria–Peloritani Terrane; Italy

Mots clés : Stratigraphie ; Conodontes ; Paléozoïque ; Unité de Stilo ; Terrane Calabro-Péloritain ; Italie

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1. Introduction

Orthoceratid limestones are typical facies of the Silurian and Early Devonian North-Gondwanic margins. In the Mediterranean Alpine Belts of southern Europe and North Africa these facies have been widely recognised: from the Pyrenees [12,28] to the Alps [29,30], from the Gibraltar Arc [23,27] to the northern Apennines [8,34], and also in Sardinia [15,16]. However, until now, such facies had never been seen in Calabria. In this article the presence of Lower to Middle Devonian orthoceratid-bearing limestones is reported for the first time in the basement of the Stilo Unit of southern Calabria. This is also the first record of Lowermost Devonian deposits, well dated by conodonts, in the Calabria–Peloritani Terrane (CPT). This opens new perspectives to the understanding of the complex Palaeozoic history of the perimediterranean Alpine areas.

2. Geological setting and previous studies on Calabrian Palaeozoic successions

The Calabria–Peloritani Arc, or Terrane [4], is an arc-shaped orogenic belt connecting the Apennine and the Maghrebian Chain. It is characterised by Alpine continental crust nappes which thrust over H-P ophiolite-bearing thrust sheets, these latter originating from the central-western Tethys. In the CPT, two sectors, characterised by different nappe architecture and Alpine tectonic evolution, have been recognised [3,4]. The southern sector of the CPT is formed only by continental crust nappes and slices, which override the Maghrebian Monte Soro Unit. The highest tectonic unit of this sector is the Stilo Unit. The nappe stack was established, and underwent a very rapid exhumation and erosion, in the Aquitanian–Early Burdigalian. This orogenic evolution is testified to the Aquitanian terrains involved in the nappe stack, and to the Burdigalian Stilo-Capo d'Orlando Formation (Fm.), sealing the nappes (Fig. 1A) [5].

The nappes are characterised by pre-Triassic basements, made up of both metamorphic and plutonic rocks, and by Meso-Cenozoic covers. In these basements, fossiliferous beds are very scarce, and the knowledge of the stratigraphic features of the successions is scanty and poorly defined. Consequently, different tectonosedimentary reconstructions have been proposed for these fragments of the Variscan orogeny involved in the Alpine Chains. In particular, the location of the CPT basements in the Variscan Chains of the central-western Mediterranean area is subject of debate [1,32].

In the 1970s, studies on the pre-Alpine basements of the CPT showed the only presence of Middle-Upper Devonian dacryoconarids and conodonts in metacalcareous lenses of both Stilo and Longi-Taormina Units [2,9,14,20]. Later, successions starting in the Cambro-Ordovician and ending with Carboniferous Culm facies were recognised in the Sila, Stilo and Longi-Taormina Units [6,7,21,22]. However, only Cambro-Ordovician and Middle-Upper Devonian layers have been dated by means of a very few specimens of acritarchs and conodonts, respectively. Consequently, the reconstructed successions and their ages are questionable because they are based mainly on facies similarities and/or field characters of volcanics occurring in them (see discussion in Spalletta and Vai [32]).

3. Stratigraphy of the Stilo Unit

The Stilo Unit is the highest thrust-sheet of the Calabrian nappe stack and crops out only in southern Calabria. It consists of both a pre-Alpine basement and a Meso-Cenozoic cover, this latter severely reduced by erosion (Fig. 1A). The basement is made up of Variscan metamorphites in which a gradual transition from amphibolite facies to greenschist facies is recognisable [3]. The low-grade metamorphites consist mainly of phyllites and metarenites, with minor metabasites, metalimestones and black cherts. The metamorphites are intruded upon by Late-Variscan calc-alkaline plutonic rocks. The sedimentary cover starts with continental redbeds a few metres thick, followed by Upper Triassic?–Lower Jurassic? dolostones and limestones. The succession continues up to the Jurassic–Cretaceous boundary with platform facies, consisting of limestones and calcareous breccias, with frequent lacunae marked by bauxitic clay layers. Upper Cretaceous–Eocene marine rocks have never been documented. The Earliest Oligocene is represented by marsh marls, marly limestones and algal limestones, followed by a layer of red clays and by Upper Rupelian–Chattian calcarenites and calcareous breccias with *Lepidocyclina*, rich in metamorphic and granitoid clasts fed by the underlying basement. The succession ends with a few metres of Aquitanian turbiditic sandstones. Alpine metamorphism is lacking, and the whole succession is unconformably covered by the Stilo-Capo d'Orlando Fm.

The Palaeozoic low-grade metamorphic succession widely outcrops in the Stilo-Bivongi area, along the Fiumara (stream) Stilaro and its tributaries, Pardalá and Crocco, where it constitutes the so-called Bivongi series [7,21], in which six main lithostratigraphic units have been distinguished. The lowest layers are represented by

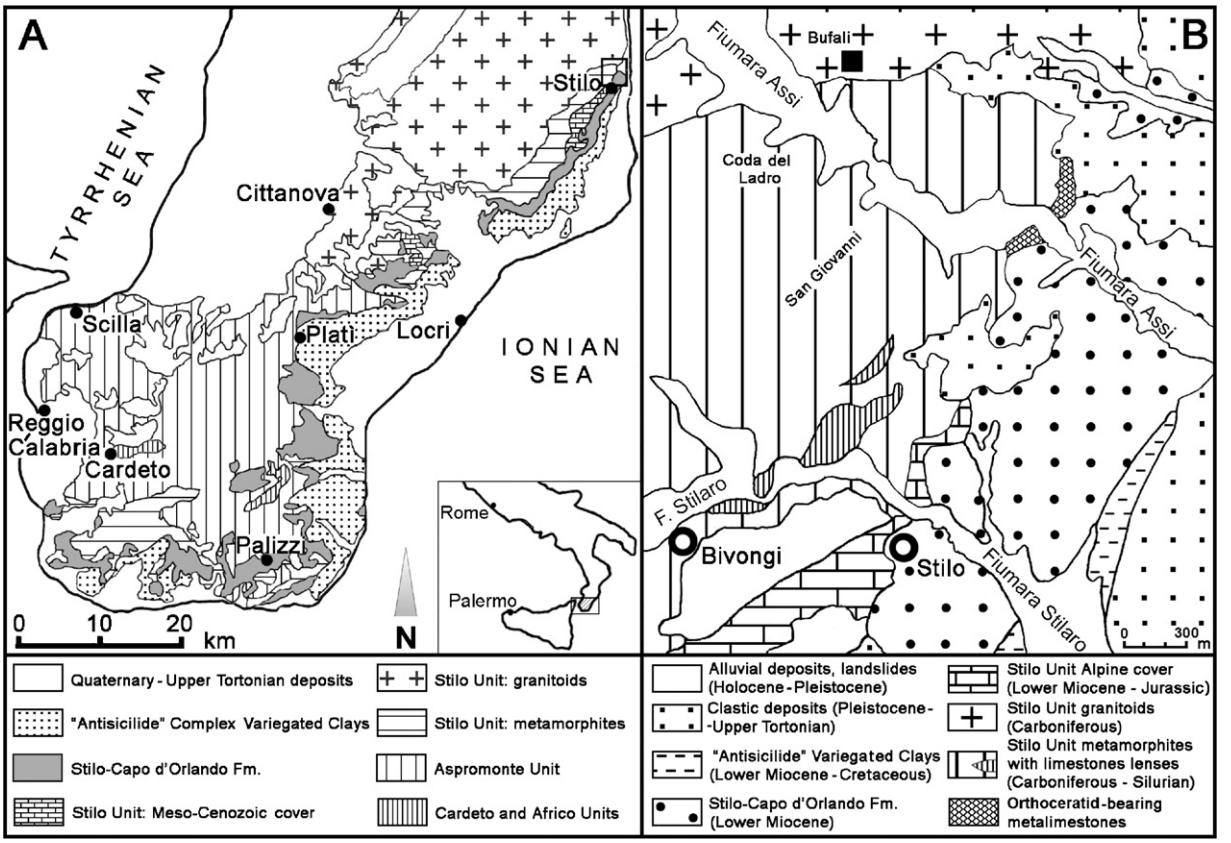

Fig. 1. **A**. Geological sketch map of the southern Calabria. **B**. Geological map of the area around the Fiumara Assi and Fiumara Stilaro.
Fig. 1. **A**. Schéma géologique de la Calabre méridionale. **B**. Carte géologique aux environs des rivières Stilaro et Assi.

several hundred metres of low-grade silvery phyllites and grey metarenites with intercalated lenses of metabasites, that Bouillin et al. [7] attributed to the Cambro-Ordovician. Its top is marked by a thin and laterally discontinuous horizon of black metapelites with some associated metalimestone lenses, with less than 10 m of total thickness and both considered to be of Silurian age. These rocks are followed by grey metapelites, alternating with thin- and fine-grained metarenite beds, more than 200 m thick, locally including greenish quartzites and, towards its top, thin-bedded limestones with some dacryoconarids, indicating a Devonian age [2]. Upwards, several strongly folded and laterally discontinuous limestone lenses, a few metres to several tens of metres thick and showing nodular structure and reddish colours (*griotte* facies), have provided Late Famennian conodonts [9,21] and constitute the only well-dated horizon, until now, in this succession. These limestones are topped by few metres of thin-bedded cherts, considered to be Early Carboniferous in age. Finally, the succession ends with more than 100 m of dark slates alternating with turbiditic sandstones, sometimes bearing plant remains and showing the classical Carboniferous Culm facies [7].

4. The Fiumara Assi section

4.1. Lithological succession

The Stilo Unit widely crops out on both sides of the Fiumara Assi (Topographic Map of Italy 1:50,000, sheet 584 Badolato; Fig. 1B), but it is particularly well exposed on the left versant (latitude N 38° 30′ 06″; longitude E 16° 28′ 57″), where the studied succession was measured (Fig. 2). Palaeozoic terrains consist of silvery metapelites and metarenites dipping towards southeast, followed by well-developed orthoceratid-bearing nodular limestones, unconformably overlain by the Stilo-Capo d'Orlando Fm. Limestones are affected by folds with hectometre wavelength and show a well-developed foliation. Structural and biostratigraphic data (see below) demonstrate that the study succession is overturned. This latter, divided in eight lithologic intervals (a to h on Fig. 2), is described below.

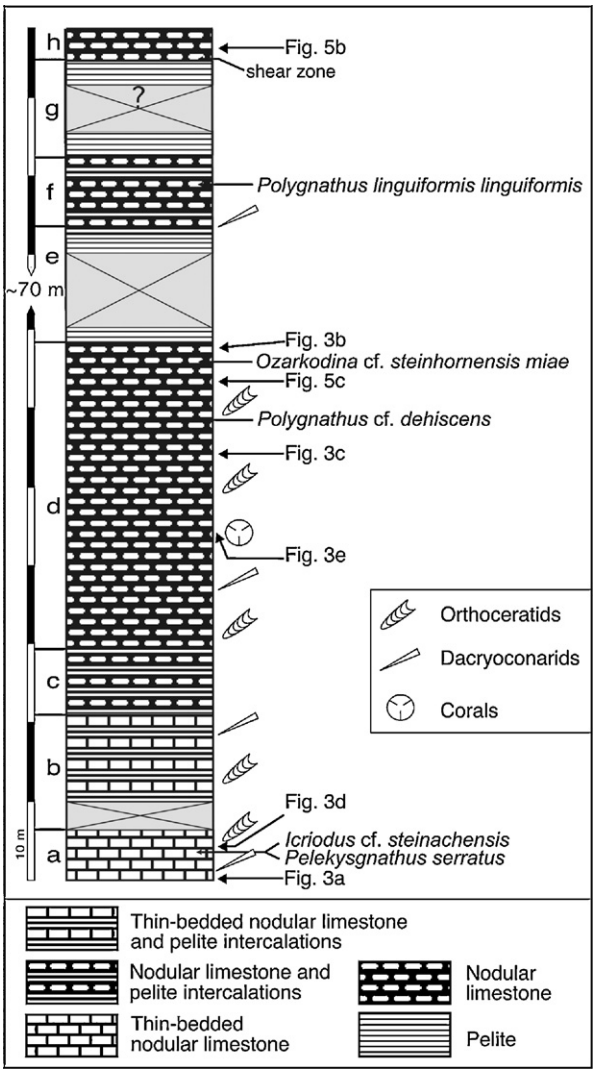

Fig. 2. Measured stratigraphic column along the Fiumara Assi. The characters a–h indicate the lithologic intervals described in the text.
Fig. 2. Colonne stratigraphique mesurée dans la Fiumara Assi. Les lettres a–h indiquent les intervalles lithologiques décrits dans le texte.

The lowermost part of the succession, lying directly under the basal conglomerates of the Stilo-Capo d'Orlando Fm. (interval a on Fig. 2; Fig. 3a), is formed by 7 m of thin- to medium-bedded, parallel-laminated and partly nodular limestones, alternating with yellowish and greyish-black fine-grained limestones, with some orthoceratids. This carbonate layer is followed by a more pelitic interval (13 m thick, partially covered at its base), formed by an alternance of thin-bedded calcareous slates, with aligned calcareous nodules, and centimetre- to decimetre-thick dark-grey limestone beds, locally with nodular structure (interval b on Fig. 2). Upwards, the succession gradually becomes more calcareous, forming an alternance (9 m) of upwardly thickening, thin- to medium-bedded limestones with progressively thinner intercalations of black pelites (interval c on Fig. 2). Interval c is followed by 40 m of limestones with very rare centimetre-thick grey to yellowish metapelites (interval d on Fig. 2). Limestones are predominantly nodular (Fig. 3b) and contain abundant orthoceratids (Fig. 3c and d) and some isolated bivalves and corals (Fig. 3e). In thin section, microfacies are essentially mudstones or wackestones with pelagic microfossils and bioclasts (dacryoconarids, orthoceratids, crinoids, and rare trilobites).

The upper part of the succession, nearly 75 m thick and partially covered, consists mainly of grey to black metapelites (intervals e and g on Fig. 2) alternating with two nodular limestone intervals (intervals f and h on Fig. 2). The limestones of the interval f (6 m) gradually change to the under- and overlying metapelites and are medium- to thin-bedded. The limestones of the interval h (Fig. 2) are strongly affected by folds with metre to decimetre wavelength, and show a sharp cleavage (see below) and a strongly sheared contact with the metapelites.

4.2. Conodont biostratigraphy

Scarce and poorly preserved conodont elements have been found in 18 of 44 limestone samples analysed (Fig. 4). Most of these can be classified only at the generic level because of strong deformation. The most abundant elements belong to *Pseudooneotodus beckmanni* Bischoff and Sannemann and *Belodella* sp. Ethington.

The oldest association (interval a on Fig. 2; Fig. 3a) is formed by *Icriodus* cf. *steinachensis* Al Rawi (Fig. 4, 1) and *Pelekysgnathus serratus* Jentzsch (Fig. 4, 4), which occurred in the delta-sulcatus Zones of the Lochkovian–Lower Pragian. Specimen 1 on Fig. 4 of *I.* cf. *steinachensis* is badly preserved but it shows a spindle-shaped outline very similar to that of the eta morphotype of the species [18]. The angle of its lateral process with the main axis of the spindle is slightly less than 90° and denticles are rounded. The different subspecies of *Pelekysgnathus* belonging to the *serratus* group are distinguished mainly on the basis of the size and position of denticles of the posterior end of blade, which are not preserved in the specimens found. Consequently, there are no indisputable criteria to assign them to any of the subspecies defined within this group.

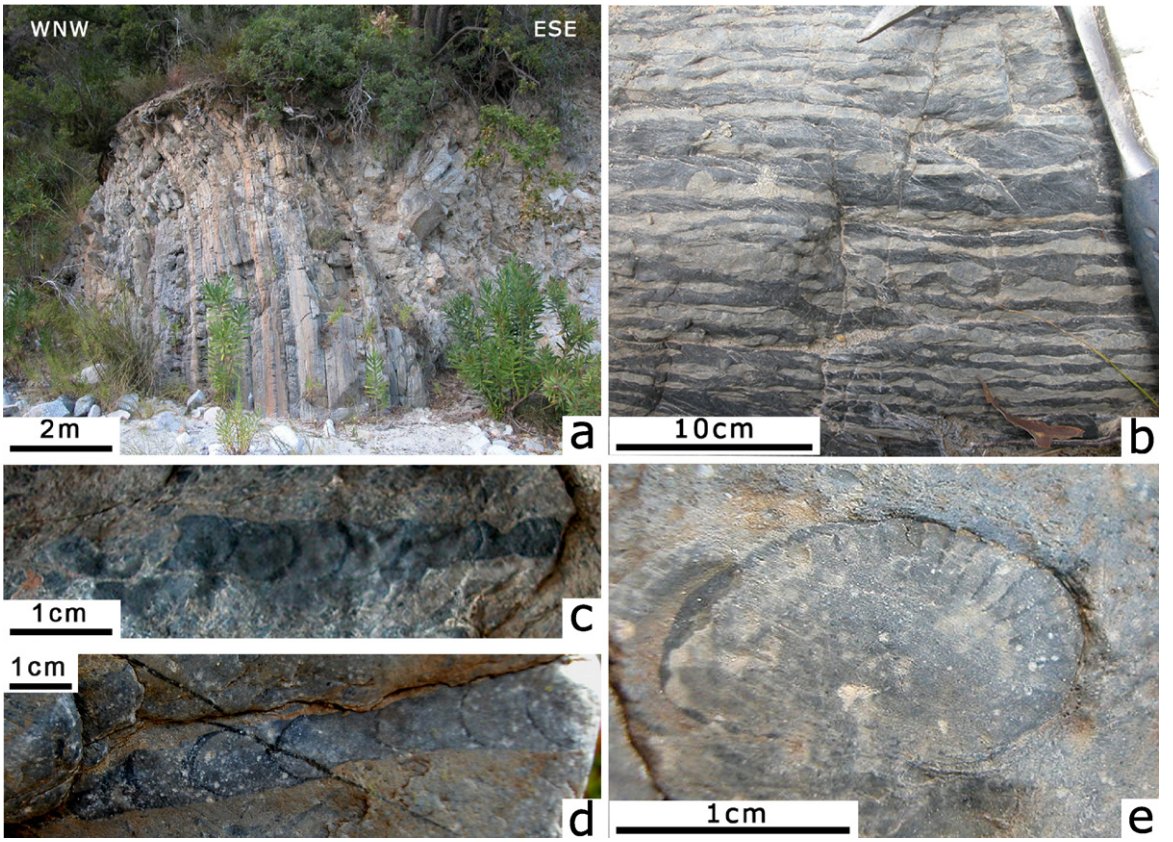

Fig. 3. **a**: unconformity of the reversed stratigraphic base of the succession below the Stilo-Capo d'Orlando Fm; **b**: nodular limestones of the upper part of the interval d; **c, d**: orthoceratid sections; **e**: flattened isolated coral. For location of photographs see Fig. 2.
Fig. 3. **a** : discordance de la succession renversée au-dessous de la Fm. Stilo-Capo d'Orlando Fm ; **b** : calcaires noduleux de la partie supérieure de l'intervalle d ; **c, d** : sections d'orthocératidés ; **e** : coraliaire isolé écrasé. Pour la localisation des photographies voir la Fig. 2.

Geometrically below the former, younger conodont associations have been found. First, several specimens of *Polygnathus* cf. *dehiscens* Philip and Jackson (Fig. 4, 3) appear in the interval d (Fig. 2). Despite its bad preservation, the specimen on Fig. 4 shows a large basal cavity and its carina does not reach the posterior end. This species belongs to the dehiscens-gronbergi Zones, indicating the topmost part of the Pragian and the Lower Emsian [25,36]. One specimen of *Ozarkodina* cf. *steinhornensis miae*, which belongs to the dehiscens-inversus Zones [18], has been found few metres upwards (interval d). Although its anterior half is broken, typical asymmetrical oval lobes are visible (Fig. 4, 2).

The youngest conodont association has been identified in the interval f (Fig. 2) located at the geometric base of the outcrop. This association is characterised by *Polygnathus linguiformis linguiformis* Hinde, a species occurring from the Eifelian to the Lower Frasnian [24].

The vertical distribution of the conodonts recorded in the limestone beds indicates that the studied stratigraphic succession has been tectonically reversed, as described below.

4.3. Structural analysis

Structural investigations of the Fiumara Assi outcrop reveal that the orthoceratid-bearing limestones are involved mainly by a mesoscale asymmetric fold system with axial planar cleavage (Fig. 5). This folding, post-Devonian in age, is responsible for deformation of the bedding (S_0), still well preserved, as well as of orthoceratids (about 50 forms/m^2) or rare corals (Fig. 3e), which appear strongly flattened on S_0 (Fig. 5c). In these folds (Fig. 5a and b), bedding (S_0) is characterised by attitudes for which the poles appear to be distributed around a large NNW–SSE circle (Fig. 5d). Folds display an axial trend moderately plunging towards northeast (Fig. 5d) and axial surfaces steeply inclined southeastwards. These northwest-verging folds are responsible for the general NE–SW strike of the bedding characterised

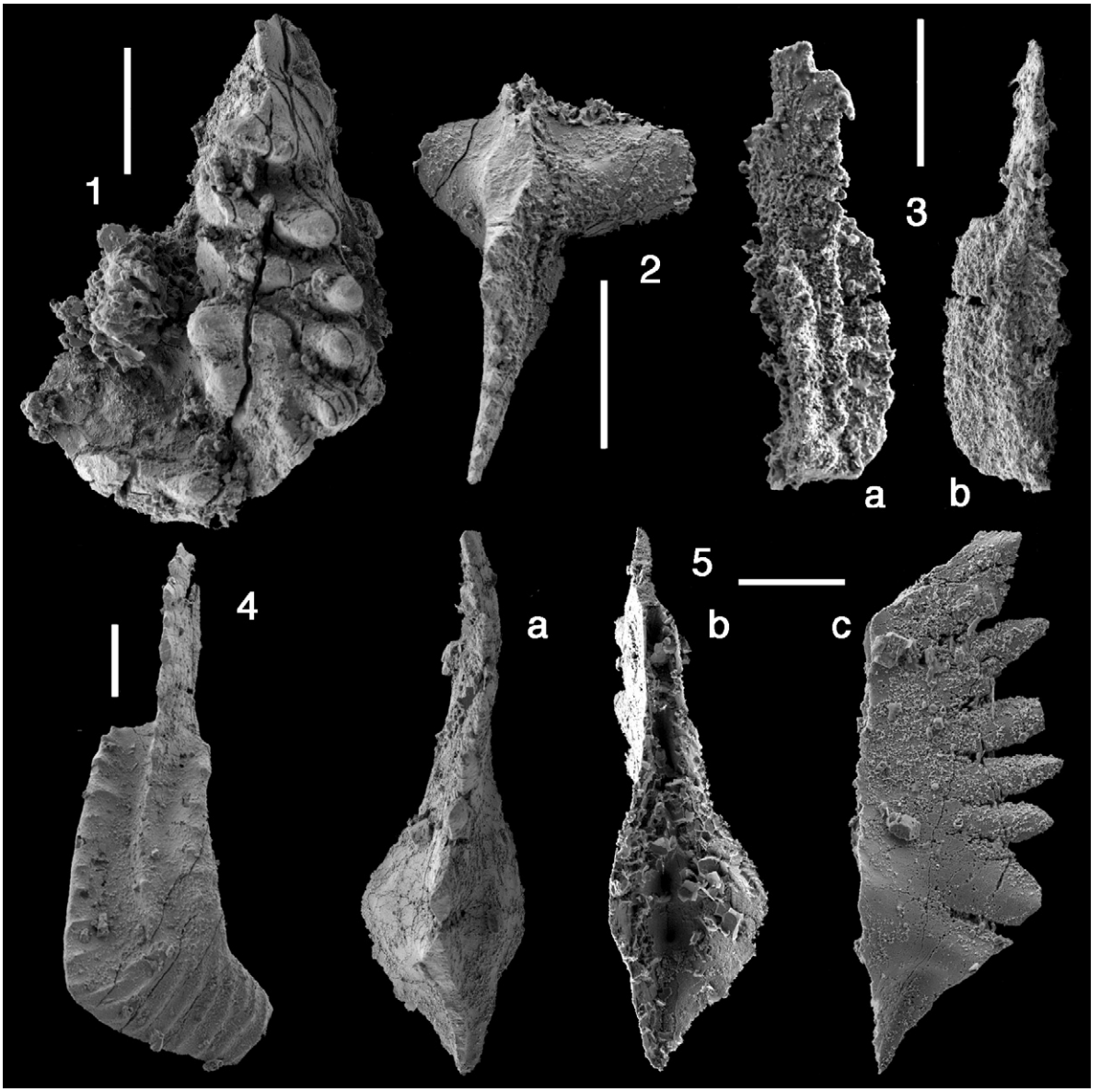

Fig. 4. **1**. *Icriodus* cf. *steinachensis* Al Rawi eta morphotype, I element (73-4). **2**. *Ozarkodina* cf. *steinhornensis miae* Bultynck, Pa element (221/86-1). **3**. *Polygnathus* cf. *dehiscens* Philip and Jackson, Pa element (82-3); **a**: upper view; **b**: lower view. **4**. *Polygnathus linguiformis linguiformis* Hinde. Pa element (216-1). **5**. *Pelekysgnathus serratus* Jentzsch, Pa element (73-1); **a**: upper view; **b**: lower view; **c**: lateral view. Scale bars: 200 μm.
Fig. 4. **1**. *Icriodus* cf. *steinachensis* Al Rawi eta morphotype, élément I (73-4). **2.** *Ozarkodina* cf. *steinhornensis miae* Bultynck, élément Pa (221/86-1). **3**. *Polygnathus* cf. *dehiscens* Philip et Jackson, élément Pa (82-3) ; **a** : vue supérieure ; **b** : vue inférieure. **4**. *Polygnathus linguiformis linguiformis* Hinde, élément Pa (216-1). **5**. *Pelekysgnathus serratus* Jentzsch, élément Pa (73-1) ; **a** : vue supérieure ; **b** : vue inférieure ; **c** : vue latérale. Échelles : 200 μm.

by vertical to steeply inclined southeast-wards attitudes, which dominate in most of the basement outcrops of the Stilo area.

The limestone beds of the Fiumara Assi are involved mainly by a northwest-verging asymmetric syncline with a hectometre wavelength. Folding is accompanied by a centimetre to millimetre spaced axial planar cleavage dipping southeastwards (Fig. 5d). The analysis of the relationships between cleavage and bedding in this syncline has allowed us to identify the normal and inverted fold limbs. In the stratigraphic succession analysed (intervals a to h on Fig. 2) S_0 is steeper than cleavage and it cross-cuts cleavage, defining angles β (*sensu* [26]) of 35–40° (Fig. 5c). Consequently, the Assi

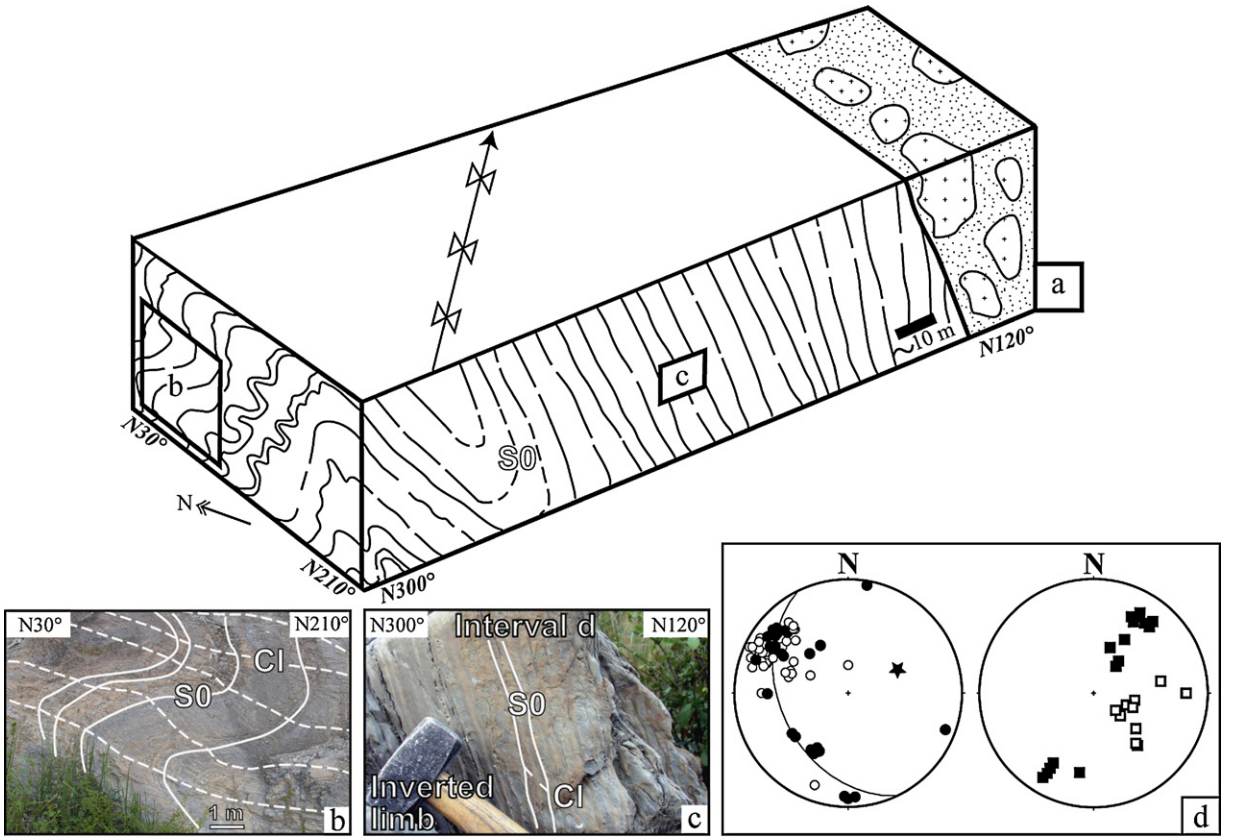

Fig. 5. **a**: block diagram showing the syncline of the Fiumara Assi outcrop (the scale bar is approximate); **b**: minor folds occurring on syncline normal limb; **c**: relationship between bedding (S_0) and cleavage (Cl) on the syncline inverted limb; **d**: orientation data (equal-area projections, lower hemisphere) of mesoscale fabrics. *Planar fabrics* (on the left): poles to bedding (S_0) distributed along π-circle of π-diagram (*sensu* [26]) (symbol: solid circle: 28 data), poles to cleavage (Cl) (symbol: empty circle: 39 data). The mean π axis is 64/50 (symbol: solid star). *Linear fabrics* (on the right): *L: S_0/Cl* intersection lineation (symbol: solid square: 19 data), movement *striae* (symbol: empty square: 14 data).

Fig. 5. **a** : schéma montrant le synclinal de l'affleurement de la Fiumara Assi (la barre d'échelle est approximative) ; **b** : plis mineurs dans le flanc normal du synclinal ; **c** : relation entre la stratification (S_0) et le clivage (Cl) dans le flanc renversé du synclinal ; **d** : stéréogrammes (projection équivalente de Schmidt, hémisphère inférieur) des fabriques à mésoéchelle. *Fabriques planaires* (à gauche) : pôles de la stratification (S_0) distribués selon un cercle π du diagramme π (*sensu* [26]) (symbole : cercle plein : 28 données), pôles du clivage (Cl) (symbole : cercle vide : 39 données). L'axe moyen est 64/50 (symbole : étoile solide). *Fabriques linéaires* (à droite) : *L : S_0/Cl* linéation d'intersection (symbole : carré solide : 19 données), stries de mouvement (symbole : carré vide : 14 données).

section constitutes the subvertical inverted limb (Fig. 5a and c) of the syncline. Along the fold normal limb, being strongly affected by cleavage and by several minor folds with metre to decimetre wavelengths, deformation has hindered a reliable stratigraphic reconstruction (Fig. 5a and b).

The intersection lineation (S_0/Cl) lies subparallel to fold hinges, and plunges both towards the northeast and southwest, defining two clusters on the equal area projection (Fig. 5d). Fine-spaced lineation plunging southeast-wards has been recognised (Fig. 5d). These linear fabrics could represent movement *striae* parallel to the slip direction associated with simple flexural slip folds (*sensu* [26]), because they lie on bedding and are arranged subperpendicular to the fold axis.

In the rare decimetre-thick metapelites, alternating with the limestones, a cleavage consisting of a fine-spaced foliation, showing sericite + quartz + chlorite ± plagioclase ± carbonate ± opaque mineral assemblage, is present. In the metapelites, P–T conditions reached the field of low metamorphism with low pressure (b_0 of the K-white micas = 8.994 [10]). In the limestones, however, the thermobaric conditions seem not to have exceeded those near the limit of the diagenesis, as still observed in the analogous limestones exposed at Pietrapennata on the southern slope of the Aspromonte Massif [14].

It is noteworthy that NE–SW trending upright folds with hectometre wavelength (up to 400 m) have been recognised along the Ionian slope of the Serre Massif,

south of Fiumara Assi, and illustrated on a geological sketch map [7]. This fold system, to which the northwest-verging folds recognised in the Fiumara Assi outcrop can be inferred, appears arranged parallel to the boundary between the Upper Carboniferous plutons and the metamorphites of the Stilo Unit exposed only a few kilometres away from the study area.

5. Concluding remarks

Lower to Middle Devonian orthoceratid-bearing nodular limestones have been firstly recognised in the CPT. These rocks belong to the Stilo Unit basement and are well exposed along the Fiumara Assi, where they are involved by an hectometre asymmetric syncline. Their age ranges between the Lochkovian–Lower Pragian and the Eifelian.

As many other Palaeozoic and younger cephalopod limestones, the orthoceratid limestones of the Fiumara Assi were deposited under pelagic conditions, with rare presence of shallow-water organisms. These conditions are typical of outer-shelf environments of moderate depth and of pelagic plateaus with condensed sedimentation [17,19], which usually occur in the distal zone of continental margins. In adjoining deeper areas, coeval Silurian–Devonian deposits were predominantly ampelitic (black shales), shaly and/or turbiditic. These are the most common lithologies of the CPT Palaeozoic successions, although they have been strongly modified by the Variscan metamorphism and orogenesis ([4], and references therein).

A similar lithological and facies spectrum is also found in Silurian–Devonian successions of other Mediterranean areas. In particular, the succession studied resembles those of SW Sardinia (Mason Porcus Fm.) [11,16,25], eastern Pyrenees [28], and other sites of the southern Variscan Chain [13]. Equivalent coeval facies are also present in the Maláguide-Ghomaride Units of the Gibraltar Arc [23,27], testifying to a common palaeogeographic origin for most of these domains. Equivalent deposits in the Alps [29,30] are interpreted as being related to the rifting and later drifting of North-Gondwanic fragments, leading finally to the opening of the Palaeotethys Ocean during Silurian–Devonian [35]. A derivation from the northern Palaeotethyan margin has been proposed for most of these western Mediterranean Palaeozoic successions [33]. In summary, the Calabrian Palaeozoic terrains and their equivalents in the western Mediterranean Alpine belts formed the western embayment of the Palaeotethys, opposite to Variscan regions of the Iberian Massif and of the Moroccan Meseta, which were rather related to the Rheic Ocean and its North-Gondwanic palaeomargin [31]. This latter was completely subducted during the Devonian–Early Carboniferous, whereas subduction of the westernmost Palaeotethys accounted later [33]. The deciphering of stratigraphic evolution of the Alpine-Mediterranean Palaeozoic successions is thus essential for understanding the Silurian–Devonian opening of the westernmost Palaeotethys and its later closing.

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SHORT COMMUNICATION

The Frasnian Upper Kellwasser event and a lower Famennian stratigraphic gap in Calabria (southern Italy)

Pilar Navas-Parejo · Rosario Rodríguez-Cañero · Roberta Somma · Agustín Martín-Algarra · Vincenzo Perrone

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Abstract The identification of three late Frasnian conodont biozones (MN 11–12 and 13, as defined in the Montagne Noire, France) and of several Famennian conodont biozones (Early, Late, and Latest *marginifera*, and Late *postera* zones of the standard zonation) in the Palaeozoic record of the Stilo Unit (southern Calabria, southern Italy) indicates the existence of a stratigraphic gap during the early Famennian (*triangularis*, *crepida* and *rhomboidea* zones). In the studied section, this gap occurs above a thin, black, pelitic horizon that appears just on top of the uppermost sample dated as late Frasnian. The conodonts obtained from the beds underlying this black horizon are all highly pyritized, unlike the few Famennian conodonts found in the limestone bed immediately overlying this horizon, which is richer in organic matter. Both stratigraphic features (the black pelitic horizon and the gap in the conodont record) are interpreted as being related to the Upper Kellwasser Anoxic Event. These new data indicate that this important biotic crisis was also evident in the pelagic environments of the Palaeozoic succession of the Stilo Unit at the very end of the Frasnian.

Keywords Upper Kellwasser · Upper Devonian · Conodont biostratigraphy · Stratigraphic gap · Calabria

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Introduction

The Late Devonian is characterized worldwide by a serious biotic crisis, generally referred to as the Lower and Upper Kellwasser (KW) Events, located close to the Frasnian/Famennian (F/F) boundary. In the lithostratigraphic record related to pelagic environments, these events are usually reflected in peculiar lithofacies such as anoxic levels (Buggisch 1972; Sandberg et al. 1988; Schindler 1990; Riquier et al. 2006; Gereke 2007; Bond and Wignall 2008) and iron or phosphate-rich surfaces related to stratigraphic gaps (Geldsetzer et al. 1993; Racki 1998; Piecha 2002). These events have been interpreted as being connected to palaeoceanographic revolutions associated with changes in the oceanographic and climatic system (e.g., Joachimski and Buggisch 1993; Becker and House 1994; Joachimski et al. 2002; Chen and Tucker 2004; Racki 2005; Riquier et al. 2006) or possibly with bolide impacts (McLaren 1970; Sandberg et al. 1988, 1997, 2002).

In the Palaeozoic successions of the Western Mediterranean Alpine orogen, Upper Devonian strata have been recognized in several isolated outcrops, but the KW Events are poorly recorded, being detected only in the Carnic Alps (Wolayer Glacier Section; Kaiser et al. 2008 and references therein) and the Malaguide Complex of the Betic Cordillera

(Rodríguez-Cañero 1993). Upper Devonian strata are also present in Variscan outcrops within other neighboring chains, which were deformed to a lesser extent during the Alpine tectogenesis (Pyrenees, Iberian Chain, and Catalonian Coastal Ranges in Spain: García-Alcalde et al. 2002; and Sardinia: Corradini 2008). In the Wolayer Glacier Section, the Lower KW horizon is represented by several centimetres of black shales. Although there are no anoxic beds at the F/F boundary, the Upper KW Event can be detected by a positive carbon isotope excursion (Joachimski and Buggisch 1993; Bond et al. 2004). In the Malaguide Complex of the Betic Cordillera, the KW Events have been found as discontinuity surfaces with Fe-rich nodules and crusts, related to very narrow stratigraphic gaps that have been dated precisely by conodonts (Rodríguez-Cañero 1993).

Until now, only Famennian beds have been found in the Stilo Unit (near Bivongi) of the Calabria-Peloritani Terrane (southern Italy, Fig. 1), dated by means of conodonts (de Capoa-Bonardi 1970). The aim of this preliminary contribution is to report the first late Frasnian conodont associations found in Calabria, which neighboring black pelites are probably coeval to the Upper KW Event. We also report new associations that permit a precise recognition of several Famennian conodont biozones.

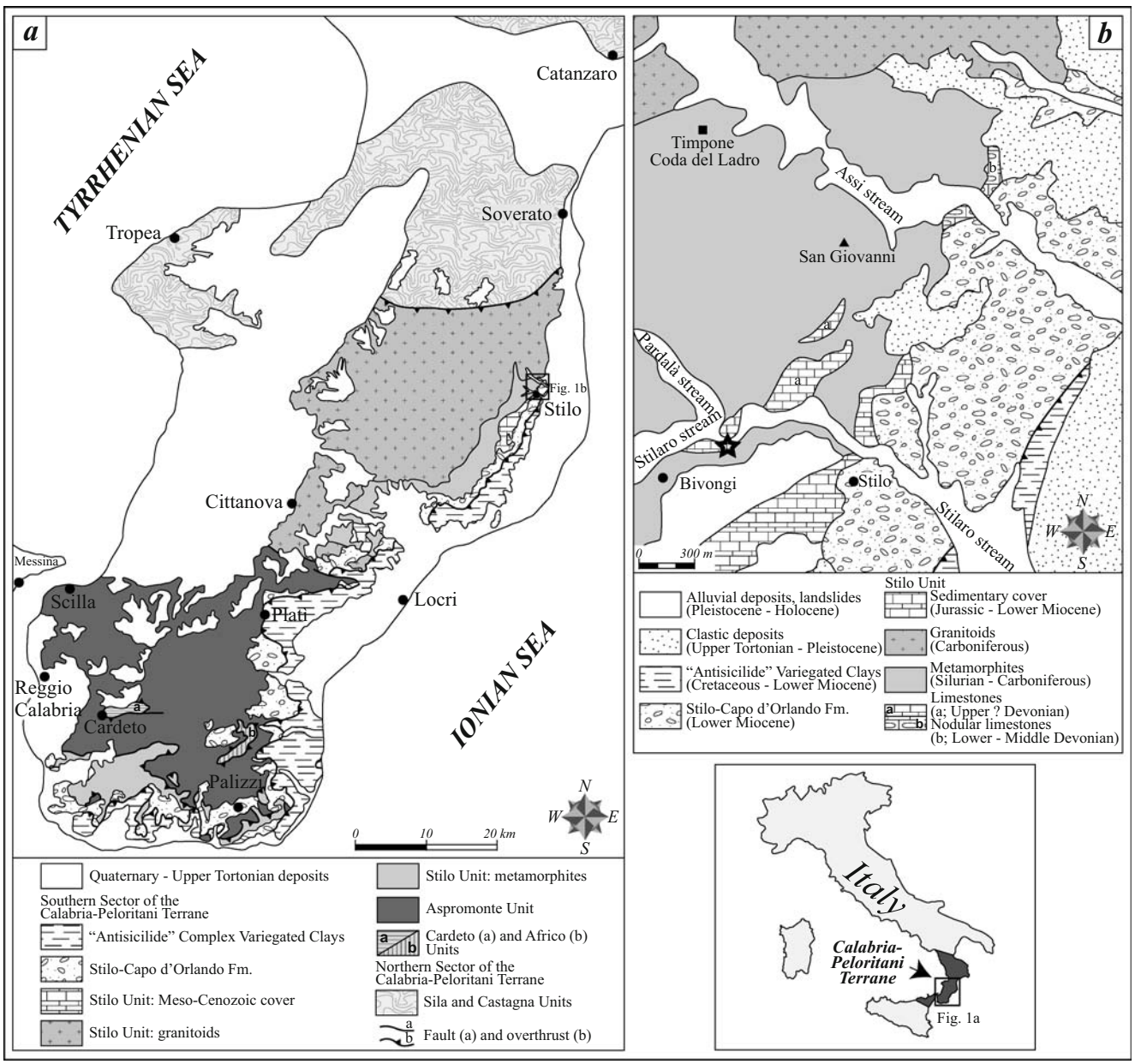

Fig. 1 **a** Geological sketch map of southern Calabria; **b** geological map of the study area (the *star* indicates the location of the section studied)

Lastly, we demonstrate for the first time in Calabria the existence of a stratigraphic gap between the upper Frasnian and the Famennian.

Geological setting and stratigraphy

The studied Palaeozoic succession belongs to the Stilo Unit (Fig. 1), the highest tectonic unit of the Internal Zones of the Calabria-Peloritani Terrane (Bonardi et al. 2002 and references therein), an arc-shaped orogenic belt connecting the Apennine and the Maghrebian Chains. This terrane includes Alpine continental-crust nappes which override H-P ophiolite-bearing thrust sheets deriving from the Central-Western Tethys. The Southern Sector of the Calabria-Peloritani Terrane is made up only of continental crust nappes and tectonic slices overriding the Monte Soro Unit, similar to the Maghrebian Flysch Basin.

The Stilo Unit is made up of a Palaeozoic metapelitic basement, several kilometers thick, intruded by Late Variscan granitoids, which is unconformably overlain by a Meso-Cenozoic sedimentary cover mainly composed of shallow-water carbonates (Amodio-Morelli et al. 1976). The metapelitic succession, with decreasing metamorphic degree upwards, includes limestone horizons bearing Palaeozoic fossils (Acquafredda et al. 1994). Lower Devonian metalimestone lenses bearing orthoceratids and conodonts represent the oldest dated layers in the Stilo basement (Navas-Parejo et al. 2009). Younger Devonian beds, formed by laterally discontinuous and usually grey limestone containing dacryoconarids (Afchain 1970) and conodonts (de Capoa-Bonardi 1970; Gelmini et al. 1979; Bouillin et al. 1987). Famennian limestones sometimes occur as reddish nodular limestones, resembling that of coeval beds with griotte lithofacies that outcrop widely in the Montagne Noire (Feist 1985) and Pyrenees (Majesté-Menjoulas et al. 1984). In the area between Stilo and Bivongi (Fig. 1b), the upper part of the basement is formed by a thin horizon of a black chert layer below a thick succession of presumably metapelites and metarenites of Carboniferous age.

The stratigraphic section analyzed in the studied area is the same as that previously dated by de Capoa-Bonardi (1970). It is located along the road (SP 95) from Bivongi to Monasterace, on a bend near the Vina bridge, downwater of the confluence of the Pardalà stream and the Stilaro stream (Fig. 1b). The 8-m-thick succession (Fig. 2) is made up of thin- to medium-bedded grey limestones with thin pelitic partings. The bedding of the slightly deformed succession is regular and shows no evidence of tectonic omission. The base of the section emerges from the alluvial sediments of the Stilaro stream and starts with a 50-cm-thick interval composed of thin-bedded grey limestones, followed by a 20-cm-thick, black clayey horizon with millimeter- to centimeter-thick intercalations of grey limestones. The thin pelitic interval is overlain by 20-cm-thick grey limestones. The uppermost part of the section is nearly 7 m thick and made up exclusively of dark-grey to bluish, thin- to medium-bedded limestones which are slightly nodular and bioturbated in places.

Conodont biostratigraphy

Frasnian

The zonation we apply to the Frasnian part of this section was proposed by Klapper (1989) for the predominantly carbonate, conodont successions in the Montagne Noire. The correlation with the Frasnian standard conodont zonation (Ziegler and Sandberg 1990) was provided by Klapper and Becker (1999). The lowermost bed of the section (sample 1, Fig. 2) contains *Ancyrodella buckeyensis* Stauffer, 1938 (Fig. 3, photos 1 and 2), which ranges from the *punctata* Zone to the Late *rhenana* Zone (equivalent to the MN 5 Zone and to the uppermost MN 12 Zone of Klapper 1989). This sample also provided two broken Pa elements, which we assign tentatively to *Palmatolepis uyenoi* Klapper, 2007 (Fig. 3, photos 3 and 4). The occurrence of this species identifies the MN 11–12 zones.

The index conodont species of the MN 13 Zone, *Palmatolepis bogartensis* (Stauffer, 1938) (Fig. 3, photo 5), appears in sample 2. The black pelite interval can be seen just above this bed. All the conodont specimens recovered from the upper Frasnian horizons lying below the black pelite interval are pyritized whilst the samples taken from the limestone bed lying immediately above it are not. Unfortunately, the few conodont fragments obtained in the overlying bed cannot be determined and so do not allow accurate dating.

Famennian

We apply the standard conodont zonation of Ziegler and Sandberg (1990) to the Famennian interval of the section. The first indisputable Famennian conodonts, obtained from the limestones overlying the black pelite bed, are included in sample 3, located about 50 cm above sample 2 (Fig. 2). These limestones have provided *Palmatolepis marginifera marginifera* Helms, 1959 (Fig. 4, photo 1). This is the defining subspecies for the base of the Early *marginifera* Zone and ranges to the Latest *marginifera* Zone. Sample 3 also contains *Polygnathus glaber medius* Helms and Wolska, 1967, which ranges from the Early to the Late *marginifera* zones. *Palmatolepis marginifera utahensis* Ziegler and Sandberg, 1984 (Fig. 4, photo 3), the index subspecies for the Late *marginifera* Zone, first occurs in

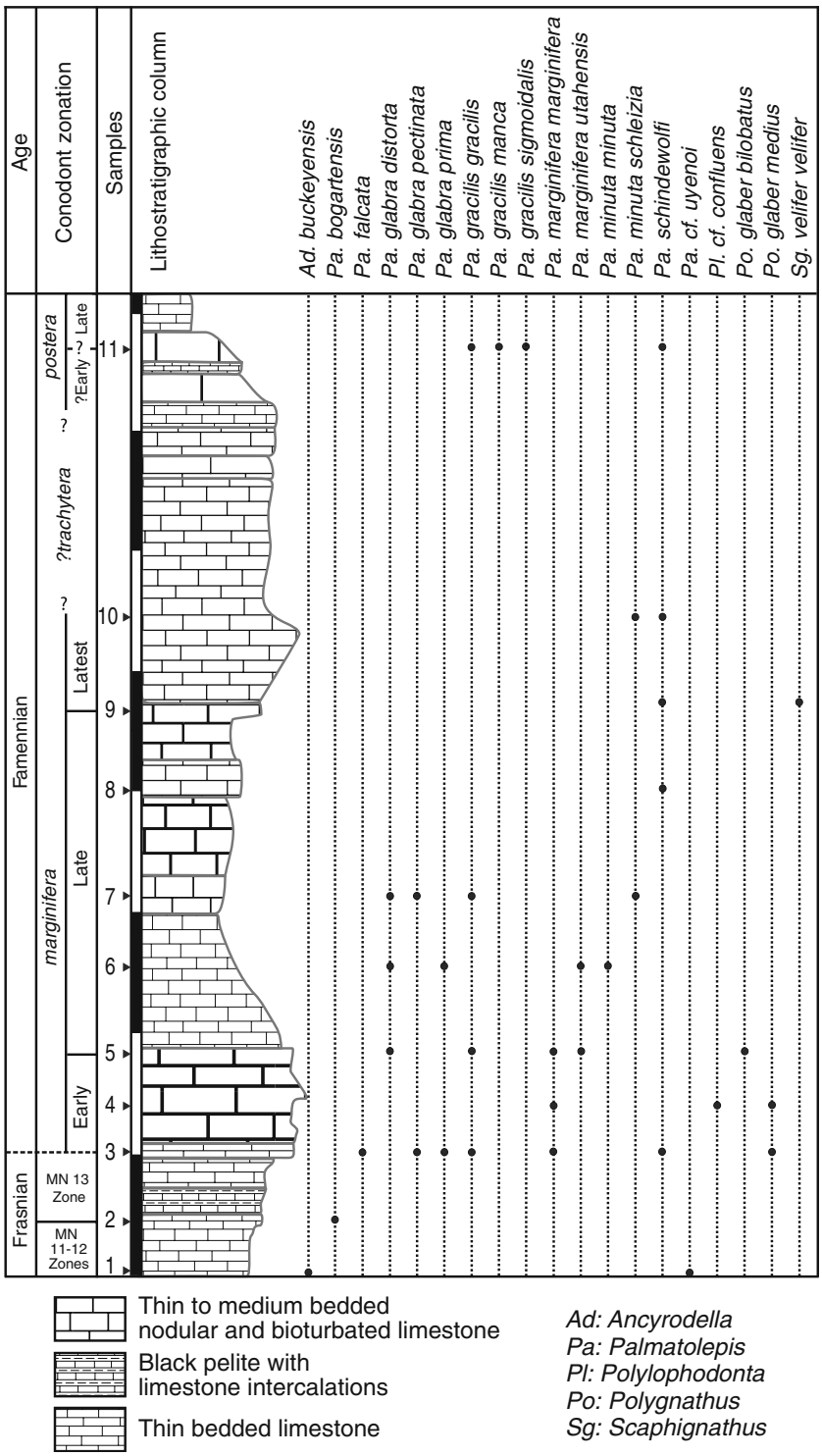

Fig. 2 Lithostratigraphic column and conodont species distribution throughout the section

sample 5. Sample 7 contains, among other species, *Palmatolepis glabra pectinata* Ziegler, 1962, whose upper range is the Late *marginifera* Zone. The occurrence of *Scaphignathus velifer velifer* Beinert, 1971 in sample 9 indicates the Latest *marginifera* Zone (Fig. 4, photo 4). We could not find any index taxa either for the *trachytera* zones or for the Early *postera* Zone in sample 10. Sample 11, however, has provided *Palmatolepis gracilis manca* Helms, 1963 (Fig. 4, photo 8), which is the key subspecies for the base of the Late *postera* Zone. Consequently the interval between samples 9 and 11 cannot currently be definitively assigned to any zone.

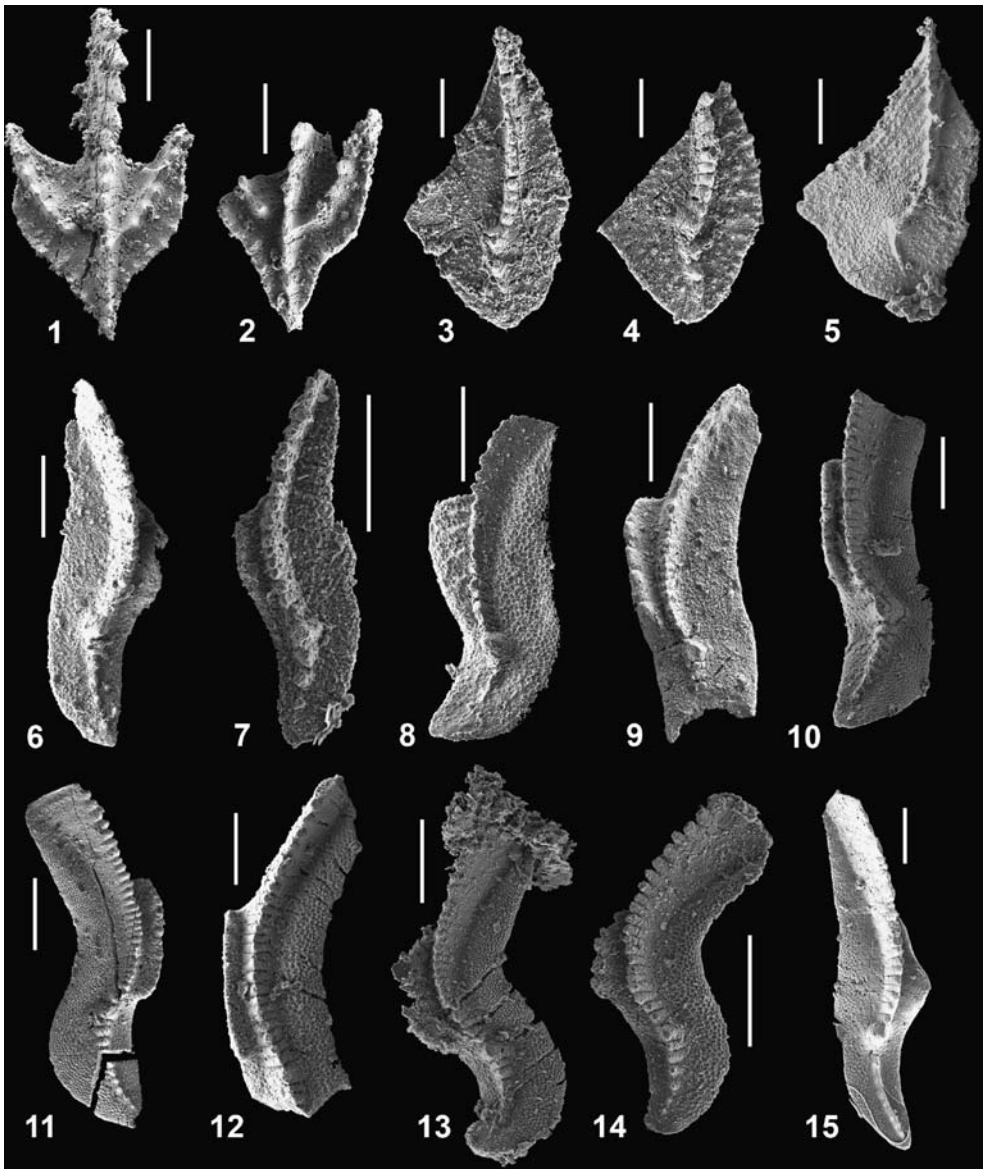

Fig. 3 All photographs are upper view of Pa elements. All conodont specimens are deposited in the Departamento de Estratigrafía y Paleontología (University of Granada). *1, 2 Ancyrodella buckeyensis* Stauffer, 1938: *1* specimen 06P-1-1; *2* specimen 06P-1-9. *3, 4 Palmatolepis cf. uyenoi* Klapper, 2007: *3* specimen 06P-1-5; *4* specimen 06P-1-6. *5 Palmatolepis bogartensis* (Stauffer, 1938): specimen 06P-2-1. *6–8 Palmatolepis glabra prima* Ziegler and Huddle, 1969: *6* specimen 06P-3-1; *7* specimen 06P-6-14; *8* specimen 06P-6-10. *9, 10 Palmatolepis glabra pectinata* Ziegler, 1962: *9* Specimen 06P-3-4; *10* Specimen 06P-7-5. *11–14 Palmatolepis glabra distorta* Branson and Mehl, 1934: *11* specimen 06P-6-6; *12* specimen 06P-5-10; *13* specimen 06P-6-13; *14* specimen 06P-7-4. *15 Palmatolepis falcata* (Helms, 1959): specimen 06P-3-15. All *scale bars* are 200 μm;

Discussion

The late Frasnian KW Crisis includes 2 of the 20 Devonian anoxic events recognized worldwide (Schindler 1993; House 2002). The Lower KW Event occurred near the base of the Late *rhenana* Zone. The Upper KW Event started during the *linguiformis* Zone and ended at or after the F/F boundary. In many stratigraphic sections, both the Lower and the Upper KW Events are observed as anoxic horizons (Sandberg et al. 1988; Paris et al. 1996; Riquier et al. 2005, 2006; Girard and Renaud 2007; Bond and Wignall 2008; among many others).

In recent years, many authors have been studying the processes, timing and causes of both KW Events. According to Girard and Renaud (2007), both KW Events were global and synchronous. The Upper KW Event coincides with a brief but widespread and intense anoxic period that is recognized worldwide usually by the presence of thin black shale horizons associated with one of the most important mass extinctions of the Phanerozoic record and with shifts in many geochemical indicators, among them stable isotopes, associated with changes in conodont biofacies (Buggisch 1972; Lottmann et al. 1986; Sandberg et al. 1988; Schindler 1990; Riquier et al. 2006; Bond and Wignall 2008; among others). In some sections of the Alpine Mediterranean region (the Wolayer Glacier Section in the Carnic Alps, in particular), the faunal extinction event is not related to the presence of any anoxic-event bed, although the marked shifts usually present in geochemical indicators have also been recognized there (Joachimski and Buggisch 1993; Kaiser et al. 2008). Most authors interpret the widespread anoxic conditions at that time as the consequence of the expansion of deep anoxic waters spreading over continental margins and platforms (McGhee 1996) due to the sinking of denser

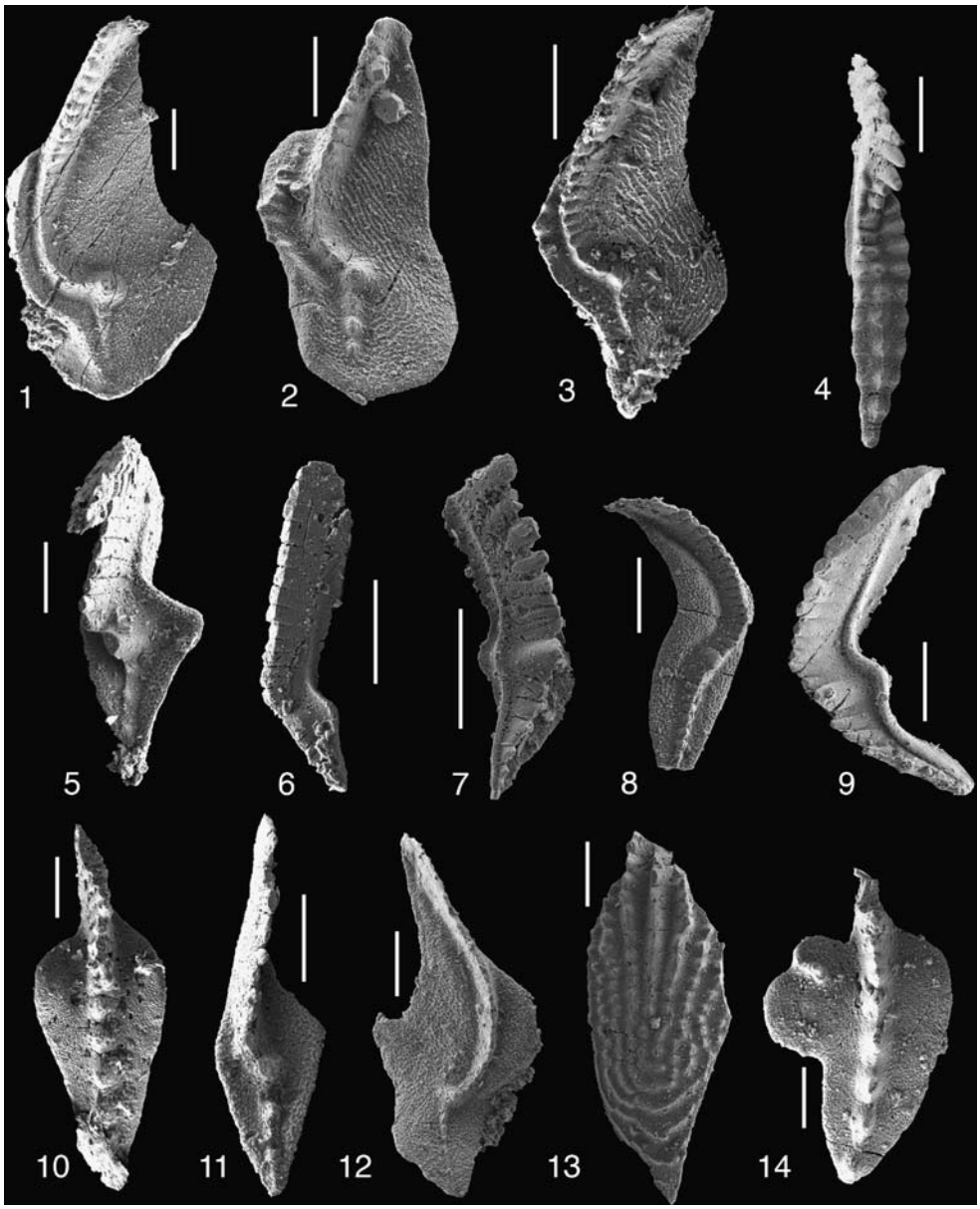

Fig. 4 All photographs are upper view of Pa elements. All conodont specimens are deposited in the Departamento de Estratigrafía y Paleontología (University of Granada). *1–2 Palmatolepis marginifera marginifera* Helms, 1959: *1* specimen 06P-3-8; *2* Specimen 06P-4-6. *3 Palmatolepis marginifera utahensis* Ziegler and Sandberg, 1984: specimen 06P-6-3. *4 Scaphignathus velifer velifer* Beinert, 1971: specimen 06P-9-3. *5 Palmatolepis minuta schleizia* Helms, 1963: specimen 06P-7-2. *6, 7 Palmatolepis gracilis gracilis* Branson and Mehl, 1934: *6* specimen 06P-5-1; *7* specimen 06P-7-6; *8 Palmatolepis gracilis manca* Helms, 1963: specimen 04A-3-16. *9 Palmatolepis gracilis sigmoidalis* Ziegler, 1962: specimen 04A-3-23. *10 Polygnathus glaber medius* Helms and Wolska, 1967: specimen 06P-3-7. *11 Palmatolepis minuta minuta* Branson and Mehl, 1934: specimen 06P-6-7. *12 Palmatolepis schindewolfi* Müller, 1956: specimen 06P-3-25. *13 Polylophodonta cf. confluens* (Ulrich and Bassler, 1926): specimen 06P-4-14. *14 Polygnathus glaber bilobatus* Ziegler, 1962: specimen 06P-5-5. All scale bars are 200 μm

and more saline waters (either colder or warmer, according to different authors: Geldsetzer et al. 1987, 1988; Wilde et al. 1988; Becker and House 1994; among others). These palaeoceanographic changes would have been triggered by a rapid fall in sea level after the Late Devonian highstand, related either to bolide impacts or to tectonic events (Sandberg et al. 1988, 2002; House 2002; Racki 1998; Averbuch et al. 2005; among others). Nevertheless, Bond et al. (2004) conclude that the Upper KW Event was an epicontinental seaway phenomenon caused by the upward expansion of anoxia from deep-basin environments. The black clay horizon is interpreted as being local evidence of anoxia in the stratigraphic record of the Stilo Unit. Nevertheless, the KW Events are not necessarily shown in the rock record at any site as anoxic (black shale) layers (e.g., Becker et al. 1991; Rodríguez-Cañero 1993; Riquier et al. 2005, 2006).

The thin black clay horizon observed in the section studied here, lying between well-dated Frasnian (sample 2) and Famennian (sample 3) beds (Fig. 2), is tentatively correlated to the Upper KW Anoxic Event. This thin black pelitic horizon appears in fact just above sample 2, which belongs to the MN 13 Zone. Its presence indicates that anoxic conditions developed in the succession belonging to the Stilo Unit at the end of the late Frasnian, whereas such conditions disappeared during the Famennian. This may explain why the Frasnian conodonts of the section studied are all pyritized below this horizon whilst those of the Famennian are usually not. Pyritization of the Frasnian conodonts during early diagenesis indicates burial under sulfidic conditions (Berner

1981) and confirms that the black pelitic horizon was deposited during the intense anoxic conditions commonly associated throughout the world with the Upper KW Event. Immediately after this anoxic event, the more or less complete reoxygenation of bottom waters would have made it more unlikely that sulfidic conditions could exist beneath the sediment–water interface, and so the conodonts deposited within the limestones postdating the anoxic event were rarely pyritized during early diagenesis associated with burial. The palaeoceanographic changes related to the development of widespread anoxic conditions might also explain why the lower Famennian *triangularis, crepida*, and *rhomboidea* zones are not recorded in the section studied. Nevertheless, the stratigraphic gap that affects the lower part of the Famennian must be confirmed after more detailed studies, as it was not possible to classify the scarce conodont fragments obtained from the few-decimetre-thick limestone beds immediately overlying the black pelitic interval. In brief, the evidence available supports the presence of the Upper KW Event beds in the Calabria-Peloritani Terrane, and this is the first record of this important event in the area.

In the Palaeozoic record of the Western Mediterranean Alpine orogen, the KW Crisis has until now only been recognized in the Carnic Alps and the Malaguide Complex of the Betic Cordillera. In this latter, an Upper Devonian conodont fauna was extracted from limestone lenses intercalated between Devonian-Carboniferous pelites and sandstones, including several Fe-rich horizons testifying to KW Events (Rodríguez-Cañero 1993). Although no KW anoxic horizon has yet been detected in the Malaguide Complex, sea-level oscillations have been detected by the analysis of conodont biofacies (Rodríguez-Cañero 1993), as it has also been described in other areas (Schindler 1993; Crick et al. 2002; Bond et al. 2004). Despite doubts concerning the significance of conodont biofacies analysis with regard to KW Events expressed in some recent papers (cf. Bond and Wignall 2008), this technique has also revealed the occurrence of oscillations in sea-level within the section studied (research in progress).

Concluding remarks

We describe for the first time the record of Frasnian beds and of a stratigraphic gap at the base of the Famennian in the Palaeozoic successions of the Calabria-Peloritani Terrane (Stilo Unit). These limestone beds, dated by means of conodonts, belong to the late Frasnian MN 11-12 and MN 13 zones, and to the Early, Late, and Latest *marginifera* and Late *postera* zones of the Famennian. The presence of a thin, black, clayey horizon rich in organic matter, located close to the F/F boundary in the Palaeozoic rocks of the Stilo Unit, could demonstrate the occurrence of the Upper KW Anoxic Event in the Calabria-Peloritani Terrane during the late Frasnian. These new findings and accurate biostratigraphic datings open possibilities for exploring in detail the incidence of the important Late Devonian palaeoceanographic events at the westernmost end of the Paleotethys.

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Variscan Tectonics in the Malaguide Complex (Betic Cordillera, Southern Spain): Stratigraphic and Structural Alpine versus Pre-Alpine Constraints from the Ardales Area (Province of Malaga). I. Stratigraphy

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ABSTRACT

In the Betic Cordillera, the pre-Alpine deformation in the Paleozoic basements is rather poorly known. At Ardales, several classic sections of the Malaguide Complex belong to two distinct, superposed Alpine tectonic units including both Paleozoic basement and Mesozoic-Tertiary cover rocks. Within the Alpine footwall unit, new stratigraphic studies allow the documentation of Upper Ordovician, Silurian, and Lower and Middle Devonian conodont associations. In particular, Silurian–Lower Devonian limestones rest on a Devonian–Lower Carboniferous succession by means of a tectonic contact, the latter sealed by unconformable Triassic redbeds. Consequently, the tectonic superposition of the studied Paleozoic successions is interpreted as the result of a Variscan thrust. The conodont color alteration index shows average values around 5–5.5, independent of the conodont association and age and the tectonic position of the studied samples. This indicates that the Malaguide Paleozoic successions were affected by a low-temperature and low-pressure Variscan metamorphism. This study provides the first solid stratigraphic evidence that documents thrust tectonics related to pre-Alpine orogeny in the Betic Cordillera. The Paleozoic stratigraphy and tectonic evolution of the Malaguide Complex are quite different from those of other Variscan regions of southern Spain and confirm its paleogeographic provenance from Paleotethyan domains located far to the east and close to the Alps, whereas most Spanish Variscan regions were connected to the Rheic Ocean, which opened to the west of Iberia and closed earlier than the Paleotethys.

Introduction

Paleozoic successions of the Betic Internal Domains (fig. 1) consist mainly of metamorphic rocks. Their general stratigraphic features and geological evolution have been reconstructed only in the less metamorphic—yet intensely deformed—rocks of the highest nappe complex of the Alpine tectonic pile: the Malaguide Complex (e.g., Martín-Algarra et al. 2004*a* and references therein). The widespread occurrence of an unconformity between Paleozoic and younger deposits of the Malaguide Complex was noticed long ago (Blumenthal 1930; Foucault and Paquet 1971). Nevertheless, relatively monotonous lithofacies, scarcity of fossils, intense deformation, and (locally) metamorphic overprint make it difficult to carry out detailed stratigraphic studies and to elucidate the pre-Alpine history of these Paleozoic rocks. In particular, the distinction between Variscan and Alpine deformation is unclear because stratigraphic data on well-dated Paleozoic successions are obtained from scattered outcrops whose correlation is difficult and because only scarce structural data have been published from them.

In this article we present a detailed stratigraphic study, together with new paleontological data, of

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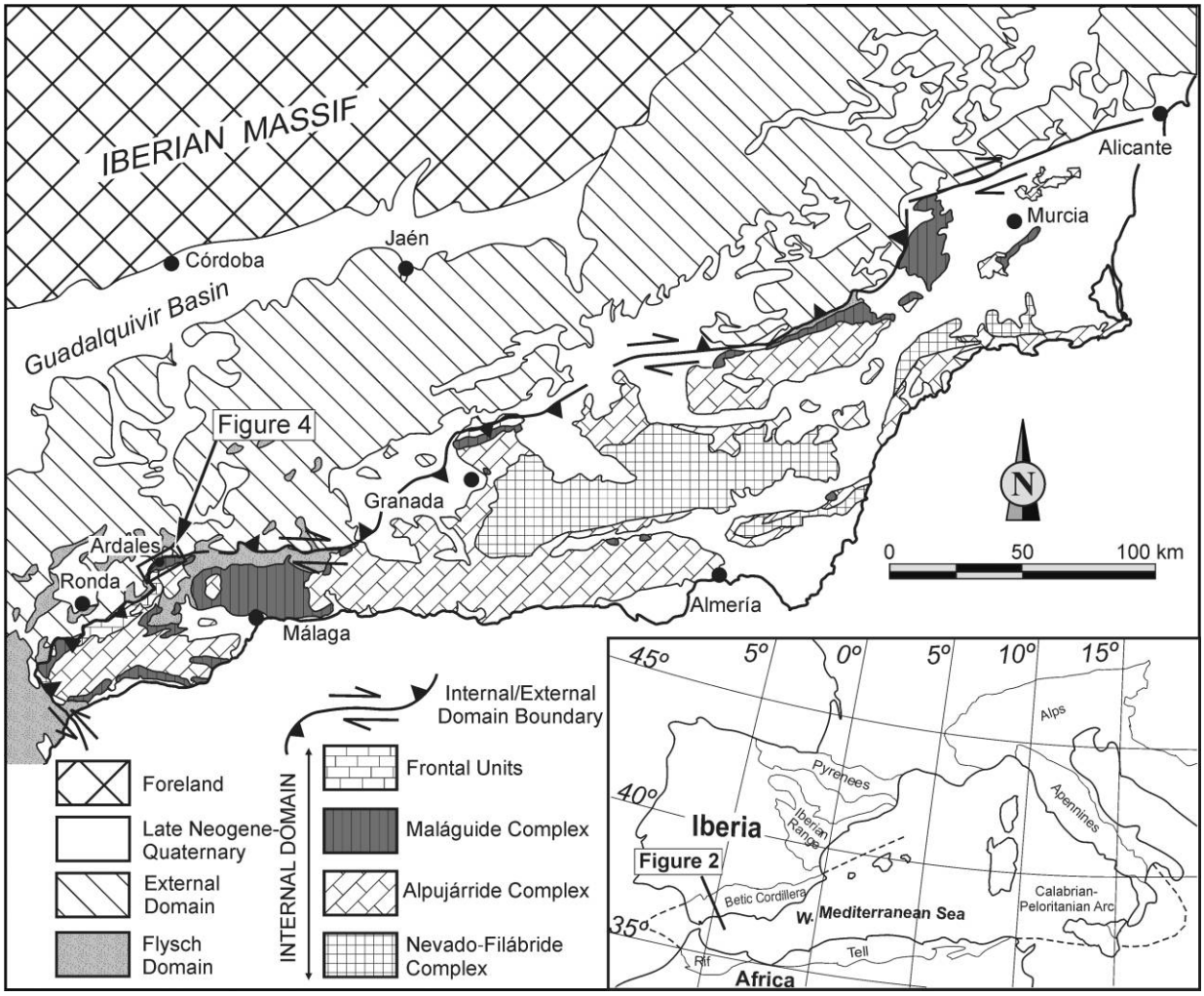


Figure 1. Tectonic sketch map of the Betic Cordillera. Modified from Vera (2004).

Paleozoic successions of classical outcrops of the Malaguide Complex in the Ardales area, in which a large amount of stratigraphic data on Paleozoic to upper Aquitanian Malaguide rocks is available (Blumenthal 1930; Kockel 1959, 1963; Kockel and Stoppel 1962; Bourgois 1978; Buchroithner et al. 1980; González-Donoso et al. 1986; Martín-Algarra 1987; Rodríguez-Cañero 1993*a*). Nevertheless, Ardales is located on the Internal-External Domain boundary, which constitutes the suture zone of the Betic Cordillera. Consequently, the area was strongly affected by the Alpine orogenic deformation, which makes it difficult to detect the Variscan or older evolution. This article demonstrates that new field and stratigraphic data make it possible to unravel pre-Alpine stratigraphic and tectonic features of the Malaguide Complex.

Geological Setting of the Western Betic Cordillera

The Betic Cordillera, the westernmost mountain belt of the Alpine-Mediterranean orogenic system (fig. 1), is divided into Internal, Flysch, and External tectonic units (Durand-Delga and Fontboté 1980; Vera and Martín-Algarra 2004). The Ardales region is located along the main contact between these major domains in the western Betic Cordillera (fig. 1). In this area a series of stacked Alpine tectonic complexes originated from the External, Flysch, or Internal domains and is exposed within a few-kilometers-wide belt. These complexes show major differences regarding their age (from pre-Silurian to late Miocene), rock type (from subcontinental mantle peridotites to sediments), and tectonic evolution.

The External and Flysch complexes consist of thin-skinned tectonic units. Those of the External Domain are made of Triassic to Neogene sedimentary successions with subordinated masses of basic volcanic and subvolcanic rocks detached from their substrata (Vera 2001; Vera et al. 2004). The basement of the External Domain consisted of southern Iberian continental crust (fig. 2). However, the basement of the Flysch Domain was made of Jurassic-Cretaceous oceanic or extremely thinned continental crust (Durand-Delga and Fontboté 1980). Deep-marine clays and turbidites, of both carbonate and siliciclastic composition (fig. 2), were deposited within this basin since the Late Jurassic up to the Early Miocene (Martín-Algarra et al. 2004*d*). By the end of the Burdigalian the cover successions of the External and Flysch domains were detached from their substratum and displaced toward the Iberian foreland (Martín-Algarra and Vera 2004*b*).

The Internal Domain forms a thrust stack of nappes, made of both continental basement rocks and Paleozoic (pre-Silurian) to Miocene sediments (Vera and Martín-Algarra 2004). Most of these units were strongly deformed and more or less affected by Alpine metamorphism during and after nappe stacking (fig. 2). The Internal Domain is usually subdivided into three nappe complexes, named, from bottom to top, Nevado-Filabride, Alpujarride, and Malaguide, plus one group of strongly imbricated frontal units (sometimes called Rondaides or Dorsale; Martín-Algarra et al. 2004*c* and references therein; Vera and Martín-Algarra 2004) outcropping along the boundary with the Flysch and External domains (fig. 1). All of these units, except for the Nevado-Filabride Complex and the lowermost Alpujarride units, occur in the western Betic Cordillera (fig. 1). In addition, subcontinental mantle peridotites (fig. 2) are widely and exclusively present within the upper units of the Alpujarride Complex near Ronda, forming the largest peridotite outcrop in the world.

Nappe stacking within the deepest part of the Internal Domain probably started during the latest Cretaceous and progressed during the Paleogene (Hebeda et al. 1980; Monié et al. 1991; de Jong 2003), up to the earliest Miocene (Martín-Algarra and Estévez 1984; Martín-Martín et al. 1997; Martín-Algarra et al. 2000, 2004*a*, 2004*c*; Martín-Martín and Martín-Algarra 2002). Nappe stacking had been completed just before erosion of the highest units of the Internal Domain started with the Burdigalian deposition of the Viñuela Group (fig. 2; Bourgois 1978; Aguado et al. 1990; Maaté et al. 1995; Martín-Algarra et al. 2004*b* and references therein), as the thrust stack was collapsing by synorogenic extension (Balanyá et al. 1997).

Outline of the Meso-Cenozoic Paleogeographic Evolution

Most of the Betic Internal Domain complexes, together with the Internal Domain complexes of the other western Mediterranean belts, formed part of a Mesomediterranean (Durand-Delga and Fontboté 1980; for a recent revision, see also Durand-Delga 2006) Triassic continental crust block (fig. 3*A*) that was separated from Iberia, Adria, and Africa during Mesozoic Pangea fragmentation and plate dispersal (Perrone et al. 2006; Critelli et al. 2008). Narrow Jurassic oceanic basins opened to the north and south of this block, separating it from the paleomargins of southern Europe–Iberia and Africa-Adria (Martín-Algarra 1987; Guerrera et al. 1993; Bill et al. 2001). The oceanic basins surrounding the newborn Mesomediterranean Microplate (MM in fig. 3*B*) joined together toward the central proto-Atlantic (CPA in fig. 3*B*), where seafloor spreading started in mid-Jurassic times (Muttoni et al. 2005).

The northern oceanic branch (fig. 3*B*) represented the western continuation of the Piedmont-Ligurian Ocean (Guerrera et al. 1993; Bill et al. 2001; Bonardi et al. 2001; Michard et al. 2002; Chalouan and Michard 2004). The high-pressure metabasites and metasediments of the Nevado-Filabride Complex (Bodinier et al. 1987; Morten et al. 1987; Puga 1990; Tendero et al. 1993; Puga et al. 1995, 1999, 2002) are ophiolitic remains of this Nevado-Filabride Ocean (NFO in fig. 3*B*).

The continental margin successions of the Betic Cordillera were deposited on both sides of this northern oceanic branch. Those of the External Domain (fig. 2) were sedimented in the southern Iberian paleomargin (SIP in fig. 3; Vera et al. 2004). Those of the frontal units, the Malaguide and Alpujarride complexes of the Internal Domain (fig. 2), also presently cropping out in the Moroccan Rif (Martín-Algarra 1987; Maaté et al. 1991, 1993; Maaté and Martín-Algarra 1992; O'Dogherty et al. 2001; Martín-Algarra et al. 2004*c*), were instead deposited on the northwestern margin of the Mesomediterranean Microplate, also named the Rondaide paleomargin (RP in fig. 3*B*; Martín-Algarra and Vera 2004*b*). In both South Iberian and Rondaide paleomargins, tectonic subsidence predominated in Triassic-Jurassic times, whereas thermal subsidence predominated during the Cretaceous. Then, pelagic conditions progressively expanded (figs. 2, 3*B*) on previous continental to shallow-

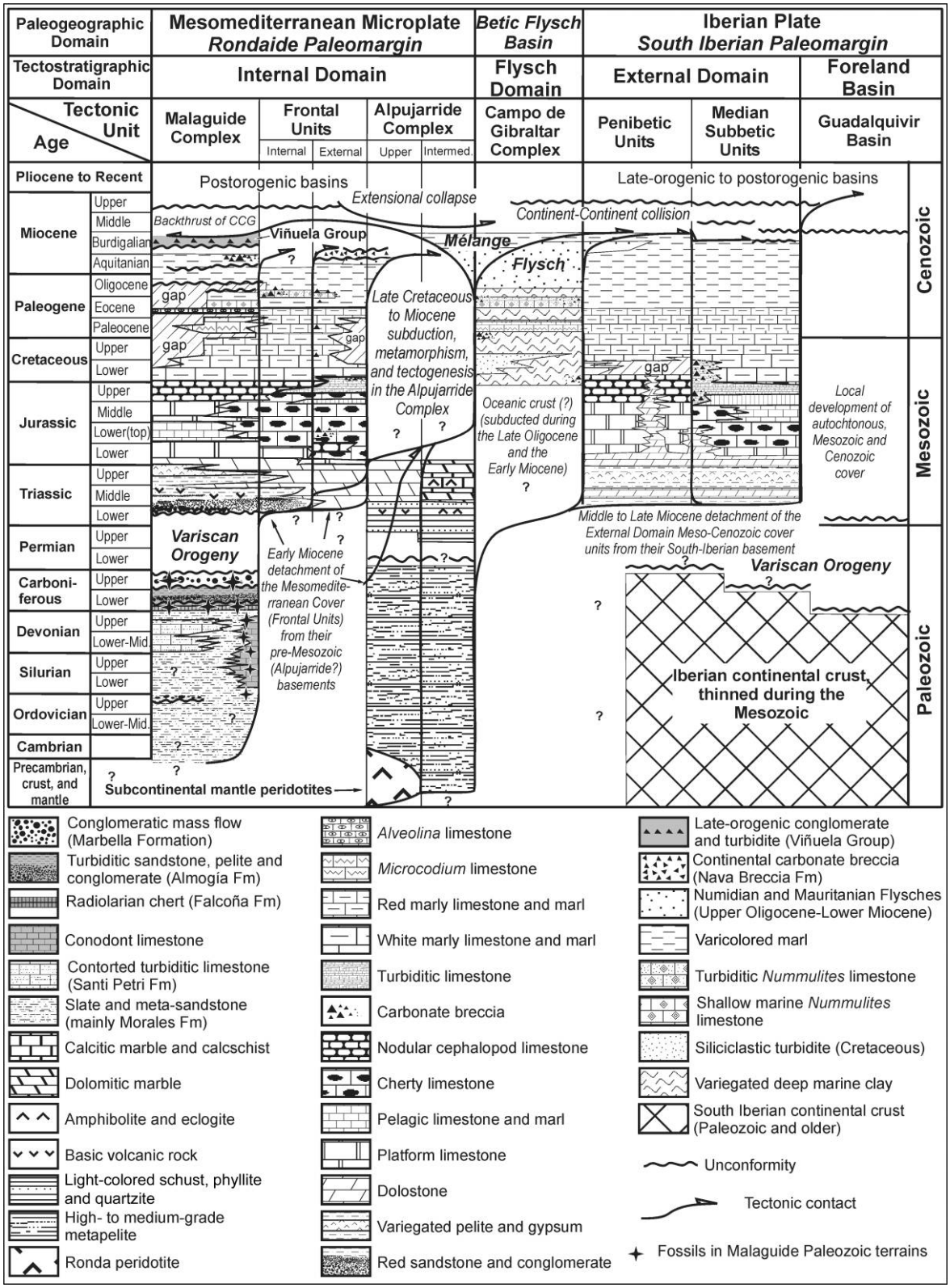


Figure 2. Very simplified stratigraphic chart of the western Betic Cordillera, with the correspondence between tectonic units and Meso-Cenozoic paleogeographic and tectonostratigraphic domains. It corresponds approximately to a NW-SE section through Ronda-Ardales-Malaga (see fig. 1).

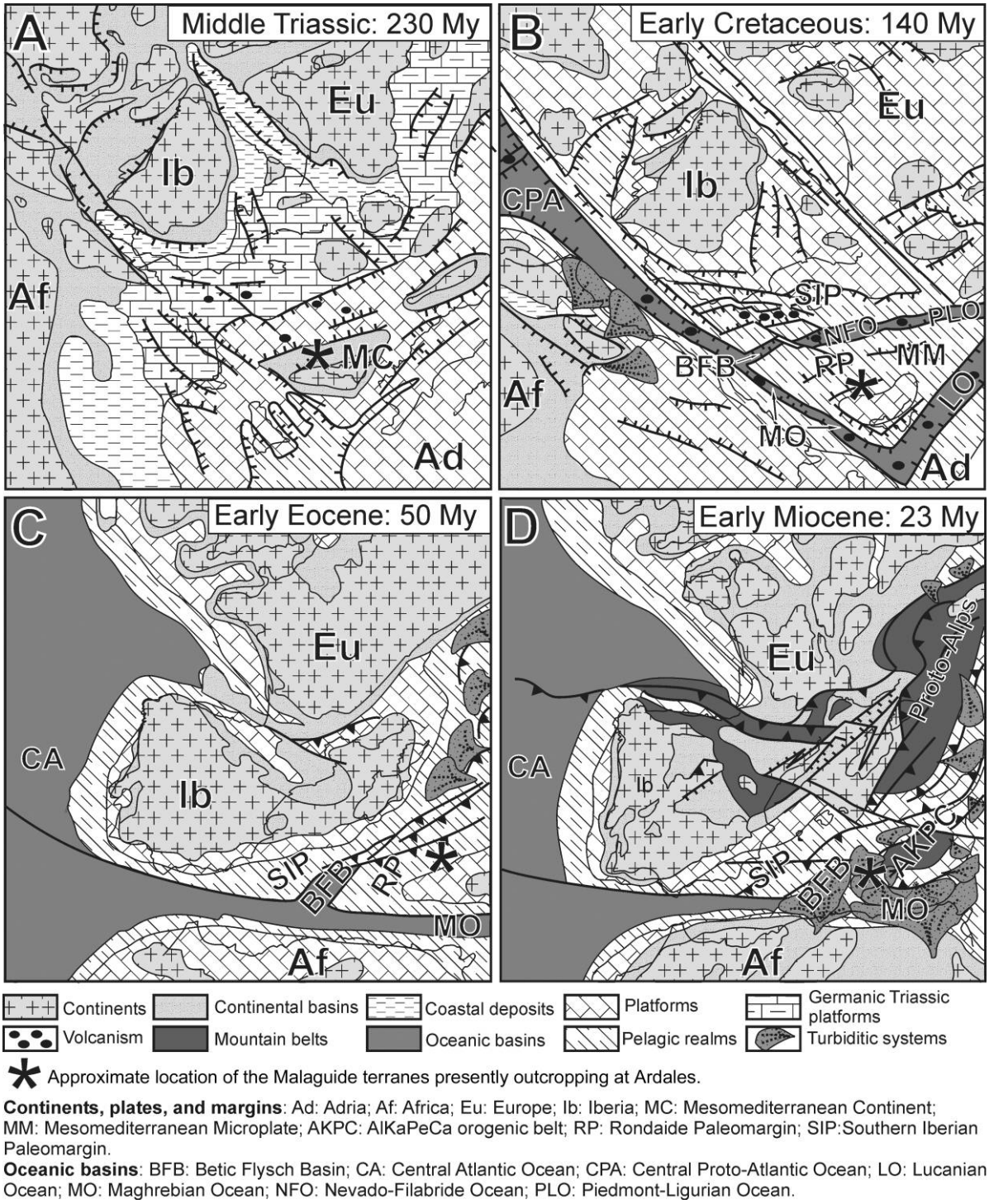


Figure 3. Paleogeographic sketches of the Betic domains during the Alpine cycle (modified from Martín-Algarra and Vera 2004*a*).

marine sedimentary environments (Vera 2001; Martín-Algarra and Vera 2004*b*).

The deep-marine sediments of the Flysch Domain units of the Betic Cordillera, included within the Campo de Gibraltar Complex, were deposited in the Betic Flysch Basin (BFB in fig. 3*B*–3*D*). The latter opened to the northwest of the Mesomediterranean Microplate in the area where the oceanic

branches that surrounded it joined together (Martín-Algarra et al. 2004*d*). The oceanic branches located to the southwest and southeast of the Mesomediterranean Microplate correspond to the Maghrebian (Durand-Delga 1980*a*; Durand-Delga and Fontboté 1980; Bouillin et al. 1986) and Lucanian oceans (Bonardi et al. 2001). The later basins (MO and LO in fig. 3*B*) opened along the northern border of Africa and along the western border of Adria, respectively, and hosted the Flysch Domain successions of the North African and Apenninic chains (Durand-Delga 1980*b*; Guerrera et al. 1993, 2005; de Capoa et al. 2003).

The ocean floor of the Piedmont-Ligurian-Nevado-Filabride Basin was subducted during Cretaceous-Paleogene convergence (fig. 3*C*). That of the Maghrebian and Betic Flysch basins was completely subducted by the end of the early Miocene. An orogenic belt named AlKaPeCa (from Alboran, Kabylias, Peloritani, and Calabria; AKPC in fig. 3*D*) originated from orogenic deformation of the Mesomediterranean Microplate (Bouillin et al. 1986; Durand-Delga 2006). After fragmentation of the AlKaPeCa and rotation of Corsica and Sardinia (Alvarez 1976), its fragments spread out to constitute the internal tectonic complexes of the western Mediterranean Alpine Belts, and the present-day western Mediterranean basins opened (Durand-Delga 2006 and references therein). The western fragment of AlKaPeCa formed the Internal Complexes of the Betic Cordillera and of the Moroccan Rif, named the Alboran Microplate (Andrieux et al. 1971; Durand-Delga 2006), Block (Martín-Algarra 1987), or Domain (Balanyá and García-Dueñas 1987). Because of the westward migration of the Alboran Block (fig. 3*D*), the oceanic (and/or thinned continental) crust basement of the Betic Flysch Basin was obliquely subducted during the Burdigalian (Guerrera et al. 1993, 2005; de Capoa et al. 2007). The leading edge of the Alboran Block acted as a southward-dipping dextral lateral ramp that accommodated its westward migration relative to the External and Flysch domains (Andrieux et al. 1971). This process led to the collision of the Internal Domain with the External Domain and to the formation of the Gibraltar Arc (Andrieux et al. 1971; Bouillin et al. 1986; Guerrera et al. 1993).

During the earliest stages of the Burdigalian collision, the contact between the Malaguide Complex and the uppermost units of the Alpujarride Complex was sealed under late-orogenic marine sediments of the Viñuela Group (fig. 2; Martín-Algarra 1987; Martín-Algarra et al. 2004*b*). This was due to extremely rapid exhumation of the upper part of the Internal Domain nappe stack (Aguado et al. 1990; Guerrera et al. 1993; Maaté et al. 1995; Serrano et al. 2006, 2007; Hlila et al. 2008). Erosion and deposition in this active tectonic setting were associated with extremely high rates of cooling and uplift of the metamorphic thrust stack in the period around 21–18 Ma (Zeck et al. 1989, 1992; Monié et al. 1994; Platt et al. 2003) and were related to continental collision (Martín-Algarra and Vera 2004*b*). Nonetheless, continental subduction toward the south and southeast of the Iberian crust (basement of the External Domain) continued in late Neogene times during persistent convergence between Africa and Iberia. This ongoing process is at present accommodated along the Internal-External Domain boundary near Granada (Morales et al. 1999).

The Malaguide Complex at Ardales

The Malaguide Complex forms the uppermost group of tectonic units of the Internal Domain (fig. 4). It is made of a strongly folded and faulted Paleozoic to Aquitanian stratigraphic succession, locally affected by low-grade metamorphism at its base (Ruiz-Cruz 2004). The Paleozoic succession (Piar Group), mainly pelitic and arenitic at the base, is followed by carbonate, cherty, and clastic sediments (fig. 5). This pre-Alpine basement is unconformably overlain by a strongly deformed Mesozoic-Cenozoic cover (fig. 5). Some geochronological data indicate the local existence of an Alpine metamorphic overprint in the basement (Platt et al. 2003).

Field data allow us to recognize in the Ardales-Turón area two Malaguide Alpine tectonic units, both well exposed and consisting of a Paleozoic basement and a Mesozoic-Cenozoic sedimentary cover (fig. 4). Actually, the Alpine age of the stacking of these two Malaguide units is demonstrated by the existence of a NE-SW-striking belt of Meso-Cenozoic rocks (fig. 4). It is made of mainly Triassic continental redbeds, structurally interposed between the Paleozoic rocks of the two Alpine units. In the field, in particular, the continental Triassic rocks and Mesozoic limestones stand out between gray to dark brown pelites and sandstones for their red color and lithology, respectively. These units (fig. 4) are hereafter named the Ardales-Mojón Unit (in the footwall) and the Castillejos-Turón Unit (in the hanging wall).

The rocks analyzed in detail in this study belong to the Alpine Ardales-Mojón Unit. This, in turn, includes two superposed subunits characterized by

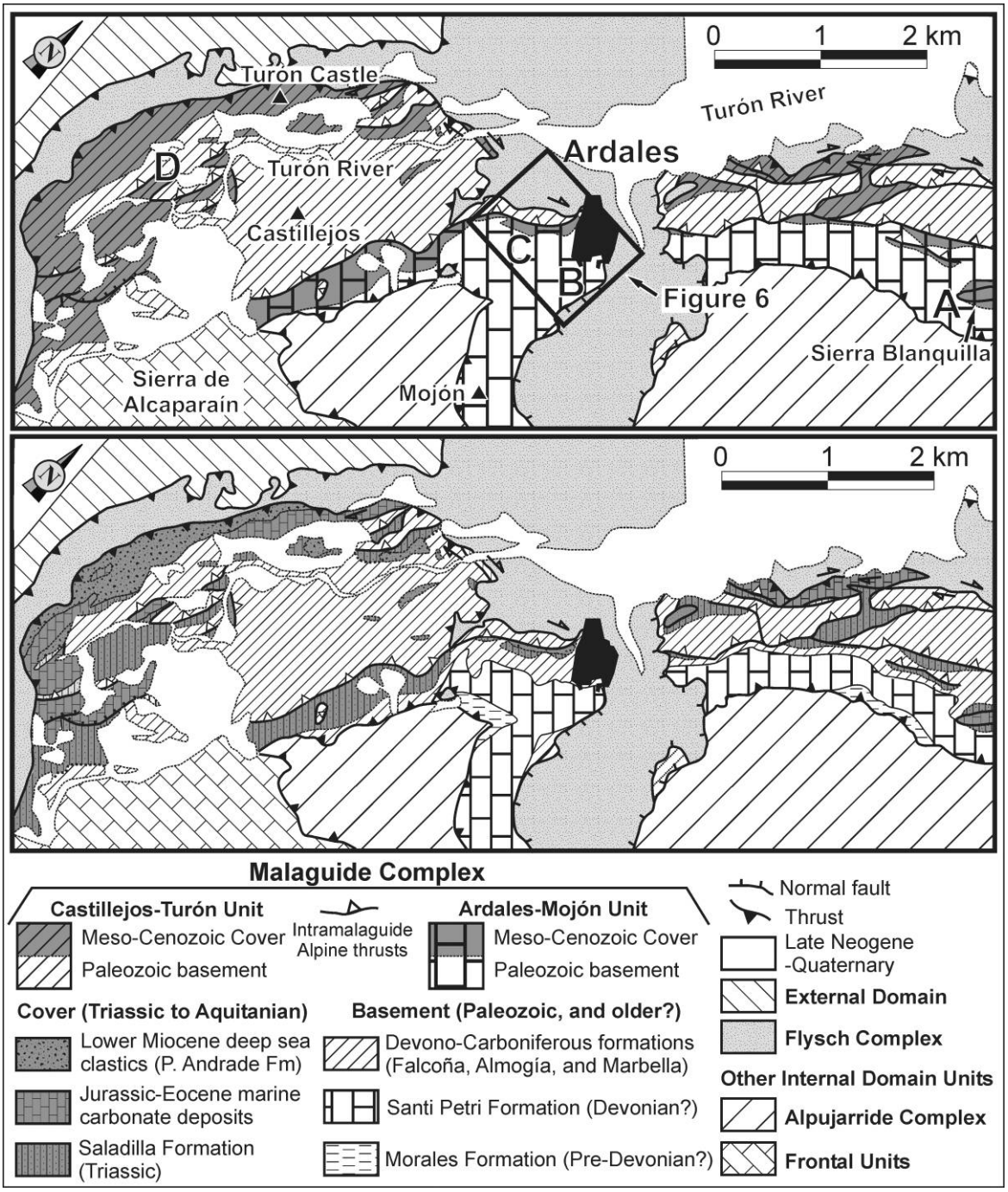


Figure 4. Tectonic map (location in fig. 1) with the main localities and simplified geological map of the Alpine units of the Malaguide Complex at Ardales. *A*, *B*, *C*, and *D* indicate the locations of the synthetic columns reproduced in figure 5.

different Paleozoic successions (both included in the Piar Group), well exposed around the Ardales Castle site (cols. B and C in fig. 5; fig. 6). In the following, we shall refer to Ermita (lower) and Peña del Castillo (upper) successions. Post-Paleozoic sediments of the Ardales-Mojón Unit also crop out near Ardales, unconformably overlying both Paleozoic successions. New age information from co-

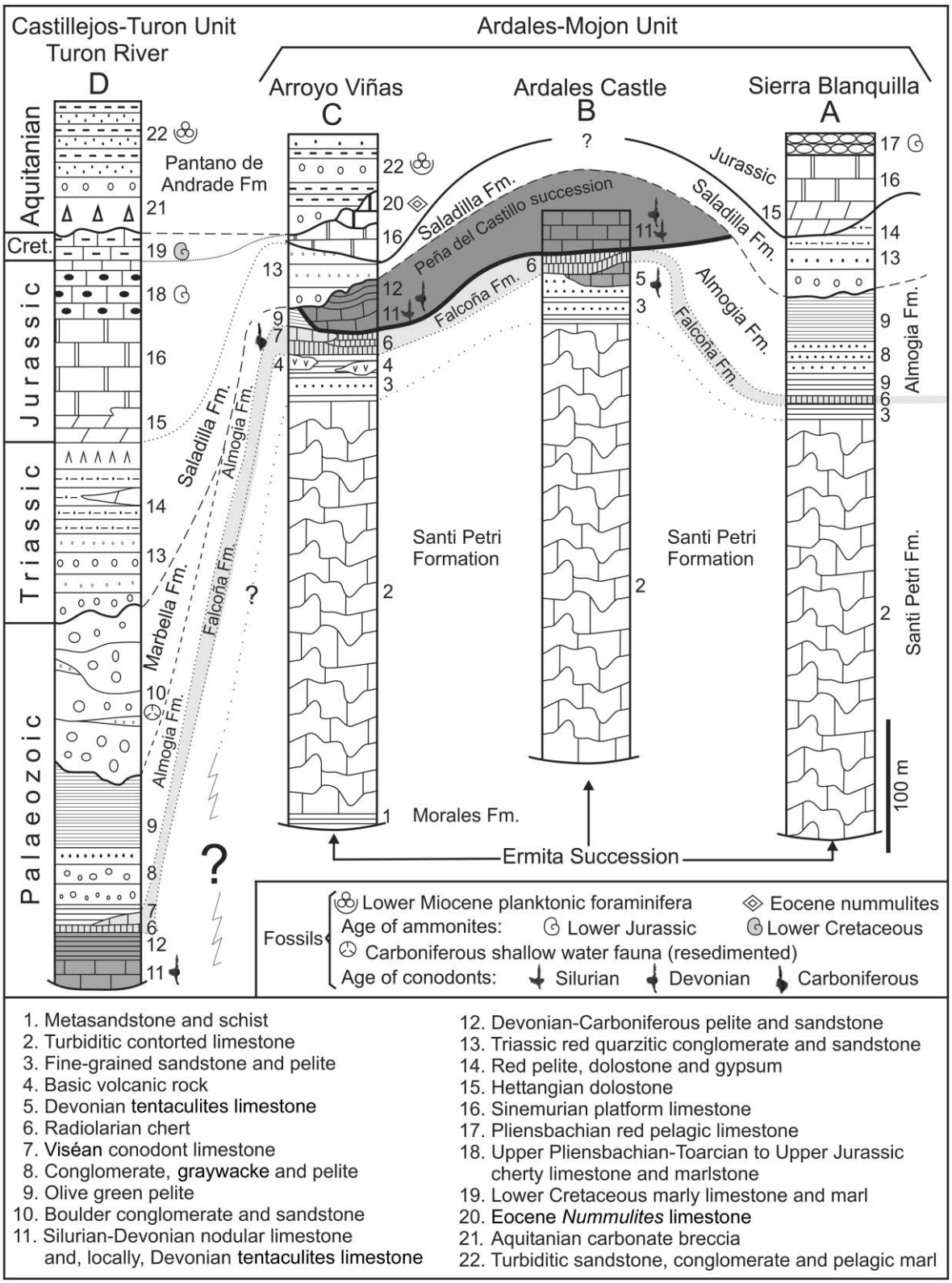


Figure 5. Synthetic stratigraphic columns of the Malaguide Complex at Ardales (locations of columns in fig. 4).

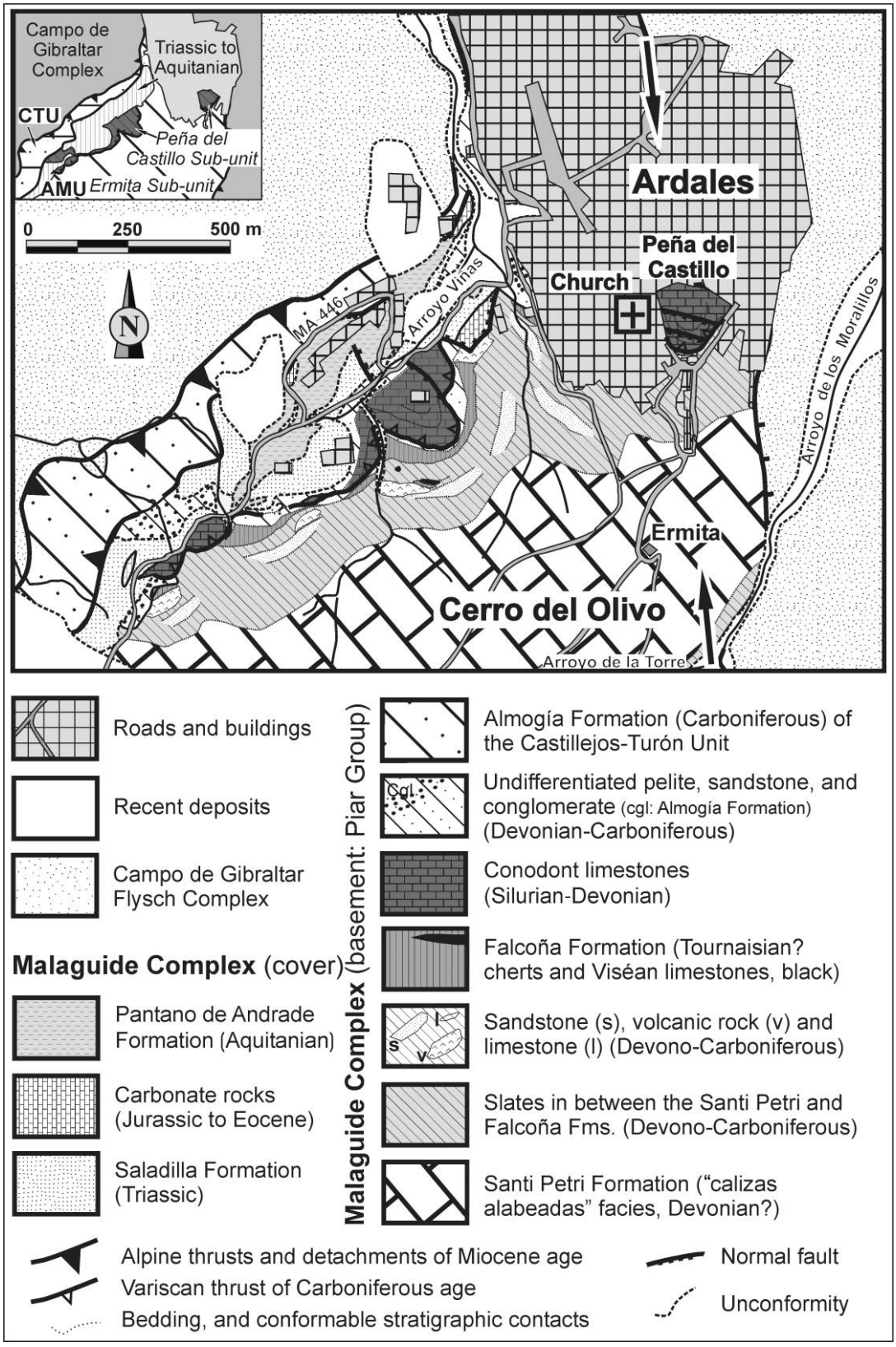


Figure 6. Detailed geological map and tectonic sketch (*inset*) of the area southwest of Ardales (location in fig. 4). *AMU*, Ardales-Mojón Unit; *CTU*, Castillejos-Turón Unit. Arrows indicate the section reproduced in figure 7*A*.

nodonts, described in this article, together with structural observations presented in a companion article (Martín-Algarra et al. 2009), demonstrates that the Peña del Castillo succession was emplaced over the Ermita succession by Variscan thrusting.

Paleozoic Successions of the Ardales-Mojón Unit

Ermita Succession. This succession crops out widely a few hundreds of meters to the northeast and southwest of Ardales (A and C in figs. 4, 5) and just south of the Ardales Castle Rock (Peña del Castillo; B in figs. 4, 5), around the Ermita del Calvario (fig. 6). It includes, from bottom to top, the Morales, Santi Petri, Falcoña, and Almogía formations. A small, lens-shaped outcrop of the Marbella Formation, with unclear (probably tectonic; Herbig 1983, p. 349) relations with underlying deposits, also exists 1.3 km to the south of Ardales. It is close to a fault between the Malaguide Complex and the Campo de Gibraltar Complex and is interpreted as a small tectonic slice of the Castillejos-Turón Unit.

Morales and Santi Petri Formations. The Morales Formation consists of a few meters of probably pre-Devonian dark gray fine-grained metapelites and metapsammites that tectonically overlie low-grade metamorphic Triassic rocks of the Alpujarride Complex. The rocks of the Morales Formation, although commonly strongly sheared, show a rapid gradual lithological transition to the overlying Santi Petri Formation by increase in the carbonate fraction. They crop out exclusively in two areas, respectively located 1 km to the southwest and 2 km to the northeast of Ardales (fig. 4; col. C in fig. 5).

The Santi Petri Formation (2, cols. A–C, in fig. 5; figs. 6, 7*A*) is several hundreds of meters thick and of probable Devonian age (Martín-Algarra et al. 2004*a*). It is made of intensely folded black to gray fine-grained turbiditic limestones and calc-schists (*calizas alabeadas* facies), associated with pelites and sandstones similar to those of the Morales Formation.

Dacryoconarid Limestone and Associated Rocks. The *calizas alabeadas* facies of the Santi Petri Formation are gradually followed by pelites and sandstones (3 in fig. 5), with lenses of basic volcanic rocks (4 in fig. 5) and of turbiditic limestones, which have never been formally included in any specifically named litostratigraphic unit (fig. 6). A Devono-Carboniferous age may be suggested for these rocks (fig. 4) because regionally they pass upward to the Lower Carboniferous radiolarites and limestones of the Falcoña Formation.

Just south of the Ardales Castle Rock, a peculiar succession may be observed in the uppermost part of the Ermita succession (3–5, col. B, in fig. 5). Here, on top of the former beds, several meters of thin-bedded, foliated, fine-grained turbiditic sandstones and pelites occur (1 in fig. 7*B*). The succession continues with calcareous shales, including tens-of-centimeter- to meter-sized lenses of pelagic limestone (2 in fig. 7*B*) that evolve upward to thin-bedded to nodular limestones with abundant dacryoconarids (tentaculitids) and conodonts (3 in fig. 7*B*). The latter form a few-meters-thick, lens-shaped body that laterally disappears and that is locally associated with carbonate breccias in its upper part as well as in its western termination (4 in fig. 7*B*). These limestones were first noticed by Blumenthal (1930) and dated with conodonts by Rodríguez-Cañero (1993*a*). Sample 187 (fig. 7*B*; table 1) provided five P_1 elements of *Ozarkodina steinhornensis miae* (Bultynck 1971) and eight ramiform elements indicating an Emsian (Early Devonian) age (11 and 12 in fig. 8). Sample 186 (fig. 7*B*) provided five small fragments of P_1 elements attributed to *Polygnathus linguiformis*? (Hinde 1879; 13 in fig. 8) and 30 elements of *Pseudooneotodus* sp. (table 1). This indicates a probable Middle Devonian age. These conodont faunas show a color alteration index (CAI) around 5.5.

A sharp surface delineates the top of the dacryoconarid limestone, and the Ermita succession above it continues with a few meters of thin-bedded, strongly deformed black radiolarian cherts (6, col. B, in fig. 5; 7 in fig. 7*B*). The lower part of these cherts (5 in fig. 7*B*) is more clayey, slightly calcareous, and rich in patinas and nodules of iron oxides. It includes one block of bioclastic sparry limestone (6 in fig. 7*B*), whose existence was previously noticed by Blumenthal (1930; K in his fig. 4). One sample from this block has provided the first Upper Ordovician conodonts found in the Betic-Rifian orogen and, up to now, the youngest Ordovician conodont association found in Spain (Rodríguez-Cañero et al. 2007). This conodont fauna shows a CAI of 4–4.5. Among other elements, this association is characterized by *Amorphognathus* cf. *ordovicicus* Branson & Mehl 1933 (14 in fig. 8) and *Scabbardella altipes* (Henningsmoen 1948; 15 in fig. 8). On the contrary, no fossils have been found in the black cherts.

Falcoña Formation. Just below the limestones of the Castle Rock, the top of the above-mentioned black cherts is strongly deformed. This contact is well observable in the field, and structural criteria certainly testify its tectonic origin (Martín-Algarra et al. 2009). Although these black cherts have not provided any fossil, they are interpreted to be part

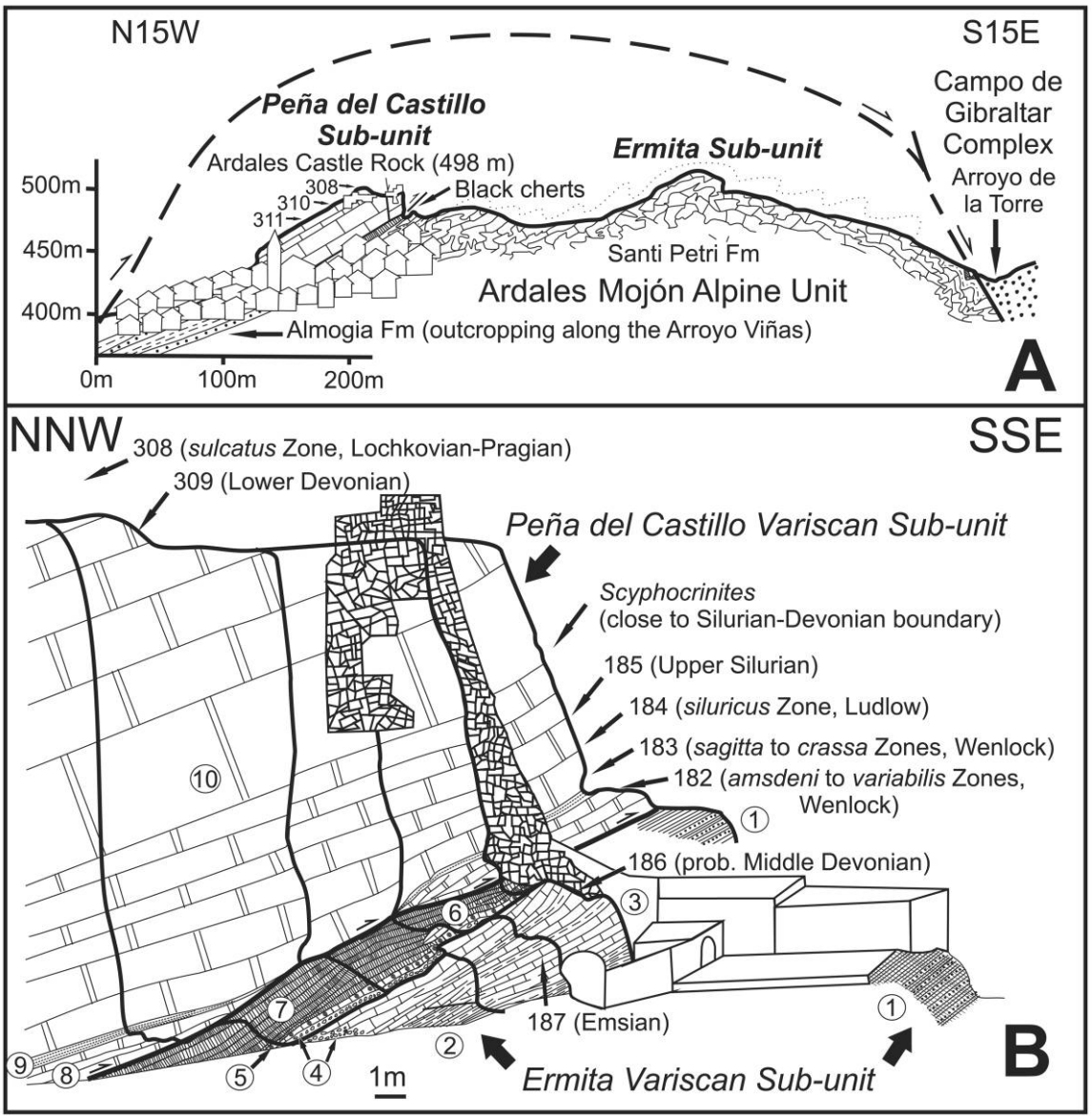


Figure 7. *A*, Geological section across the Ardales Castle site. This profile reproduces a natural panorama of the outcrop as it is visible from the west. The houses of the village and the outcrop of the Almogía Formation along the Arroyo Viñas have been projected on the section (location in fig. 6). The numbers 310 and 311 indicate the locations of the conodont-bearing samples mentioned in the text (see also table 1). *B*, Outcrop sketch (stacked series of parallel profiles marked by thicker lines) of the southern cliff of the Ardales Castle Rock, with the location of the samples dated with conodonts. *1*, Strongly foliated, alternating fine-grained sandstones and pelites. *2*, Olive-green pelites, including thin limestone beds and clasts and brecciated limestone lenses. *3*, Limestones with tentaculitids. *4*, Calcareous breccia. *5*, Nodules and patinas of iron oxyhydroxides included within slightly calcareous black cherts. *6*, Finely crystalline limestone block included in siliceous pelites, containing Late Ordovician conodonts. *7*, Thin-bedded, flattened, and sheared black cherts. *8*, Medium-bedded, locally strongly brecciated, Silurian nodular limestones with conodonts (see table 1). *9*, Thin intercalation of black slightly siliceous shales. *10*, Nodular and massive, thick- to medium-bedded Silurian limestones.

Table 1. Silurian-Devonian Conodonts from Paleozoic Successions Exposed at Ardales

	Ermita succession		Peña del Castillo succession							
Variscan subunits	187	186	182	183	184	185	311	310	309	308
Dissolved weight (kg)	.29	.34	.30	.30	.32	.31	.75	.75	.75	.75
Total no. elements	13	52	1	15	23	15	104	20	177	11
Belodella sp.		1								
Pseudooneotodus sp. B		30								
Polygnathus linguiformis? P_1		5								
Dapsilodus obliquicostatus S_2-S_3			1	2						
Pseudooneotodus beckmanni				1						
Kockelella manniki P_1					2					
K. manniki P_2					3					
K. manniki M					2					
K. manniki S_1					1					
Ozarkodina steinhornensis miae P_1	5									
Ozarkodina sagitta sagitta P_1				4						
O. sagitta sagitta P_2				5						
O. sagitta sagitta M				1						
O. sagitta sagitta S_3				2						
Ozarkodina remscheidensis remscheidensis P_1							18			
O. remscheidensis remscheidensis P_2							4			
Ozarkodina excavata excavata P_1						3	2	2	24	
O. excavata excavata P_2									3	
O. excavata excavata S_1						1	1	1	5	
O. excavata excavata M							1	1	2	
O. excavata excavata S_2						2	1	3	8	
O. excavata excavata S_3								1	4	
Ozarkodina pandora P_1										4
Ramiforms and undetermined fragments	8	16			15	9	77	12	131	7

Note. Sample numbers are shown below each succession.

of the Carboniferous Falcoña Formation (6, col. B, in fig. 5).

Few hundreds of meters to the west of Ardales, a laterally discontinuous, similar succession of cherts has been recognized and mapped (fig. 6). In detail, just to the northwest of the Ermita, in the Arroyo de las Viñas area, the following succession on top of the Santi Petri Formation has been recognized, from bottom to top (3–7, col. C, in fig. 5): (i) graywackes and pelites, (ii) discontinuous lenses of basic volcanic rocks, (iii) black cherts, and, locally, (iv) gray limestone lenses containing Viséan conodonts (*Gnathodus* cf. *bilineatus*). This is the first report of Carboniferous conodonts at Ardales. The existence of this Viséan limestone horizon confirms, biostratigraphically, the previously supposed presence of the Falcoña Formation in the area and supports the Carboniferous age of the cherts on top of the Ermita succession in the southern slope of the Ardales Castle Rock.

Almogía Formation. The Falcoña Formation is overlain by turbiditic conglomerates, graywackes, and olive-green pelites (8 and 9 in fig. 5), which constitute the Almogía Formation. The latter represents the post-Viséan "Culm" facies Carboniferous deposits of the Malaguide Complex (Martín-Algarra et al. 2004*a*). In the Ardales-Mojón Unit (Ermita succession), these rocks are better exposed to the northeast of Ardales (col. A in fig. 5), although they are also locally present to the southwest of the village (col. C in fig. 5; figs. 6, 7).

Peña del Castillo Succession. This succession is exposed at the Ardales Castle Rock (Peña del Castillo; fig. 7*A*) and in several outcrops southwest of Ardales (fig. 6). It is formed by several tens of meters of well-bedded, NNW-dipping, reddish to grayish, thick- to medium-bedded nodular ("*griotte*-like" facies) and micritic limestones (11, cols. B and C, in fig. 5) and constitutes the best-dated Silurian–Lower Devonian succession of the Malaguide Complex (Rodríguez-Cañero 1993*a*). In particular, the succession starts with medium-bedded (10–20 cm), locally strongly brecciated, brown to reddish nodular limestones (8 in fig. 7*B*). It is followed by a thin intercalation (50–60 cm) of black slightly siliceous shales (9 in fig. 7*B*). Finally, most of the Castle Rock is formed by nodular and massive, thick- to medium-bedded limestones (10 in fig. 7*B*). These limestones contain orthoceratids (Blumenthal 1930), one bed contains *Scyphocrinites* (Rodríguez-Cañero et al. 1997; see fig. 7*B*), and they provide abundant Silurian–Lower Devonian conodonts

(fig. 8). Conodont associations are reported in table 1.

Samples 182 and 183 (see location in fig. 7*B*), containing *Dapsilodus obliquicostatus* Branson & Mehl 1933 (1 in fig. 8) and *Ozarkodina sagitta sagitta* (Walliser 1964; 2 in fig. 8), represent the *amsdeni-ranuliformis* and *sagitta* zones (Wenlock). Sample 184 (fig. 7*B*), with *Kockelella manniki* Serpagli & Corradini 1998 (3 in fig. 8), falls into the *siluricus* zone (Ludlow). Samples 185, 309 (location in fig. 7*B*), 310, and 311 (location in fig. 7*A*), with *Ozarkodina excavata excavata* (Branson & Mehl 1933; table 1; 4–8 in fig. 8) and with *Ozarkodina remscheidensis remscheidensis* (Ziegler 1960; only present in sample 311; table 1; 9 in fig. 8), belong to the upper Ludlow-Pridoli-Lochkovian interval (Upper Silurian–Lowermost Devonian). Finally, sample 308 (location in fig. 7), with elements P_1 from the morphotype ε of *Ozarkodina pandora* Murphy, Matti & Walliser 1981 (table 1; 10 in fig. 8), is attributed to the *sulcatus* zone (Lochkovian-Pragian; Lower Devonian). All of these conodont faunas show CAI around 5.5.

The relationships of the Silurian-Devonian succession of the Peña del Castillo with the overlying Paleozoic and younger deposits are not visible because they are hidden by the houses of the village (figs. 3, 5). Nevertheless, these relationships may be observed a few hundreds of meters to the west of Ardales (fig. 6). In this sector, Silurian and Devonian nodular limestones, sometimes bearing tentaculites and lithologically quite similar to those of the Peña del Castillo (11, col. C, in fig. 5; Arroyo Viñas) are locally covered (fig. 6) by undated (Devono-Carboniferous?) pelites and fine-grained sandstones (12 in fig. 5) and by unconformable Triassic (13 in fig. 5) and younger deposits. The Silurian and Devonian limestones, well dated by conodonts (Kockel 1959; Rodríguez-Cañero 1993*a*), represent the western continuation of the Castle Rock succession and also overlie the Falcoña Formation cherts and associated rocks (6 and 7 in fig. 5) at the top of the Ermita succession.

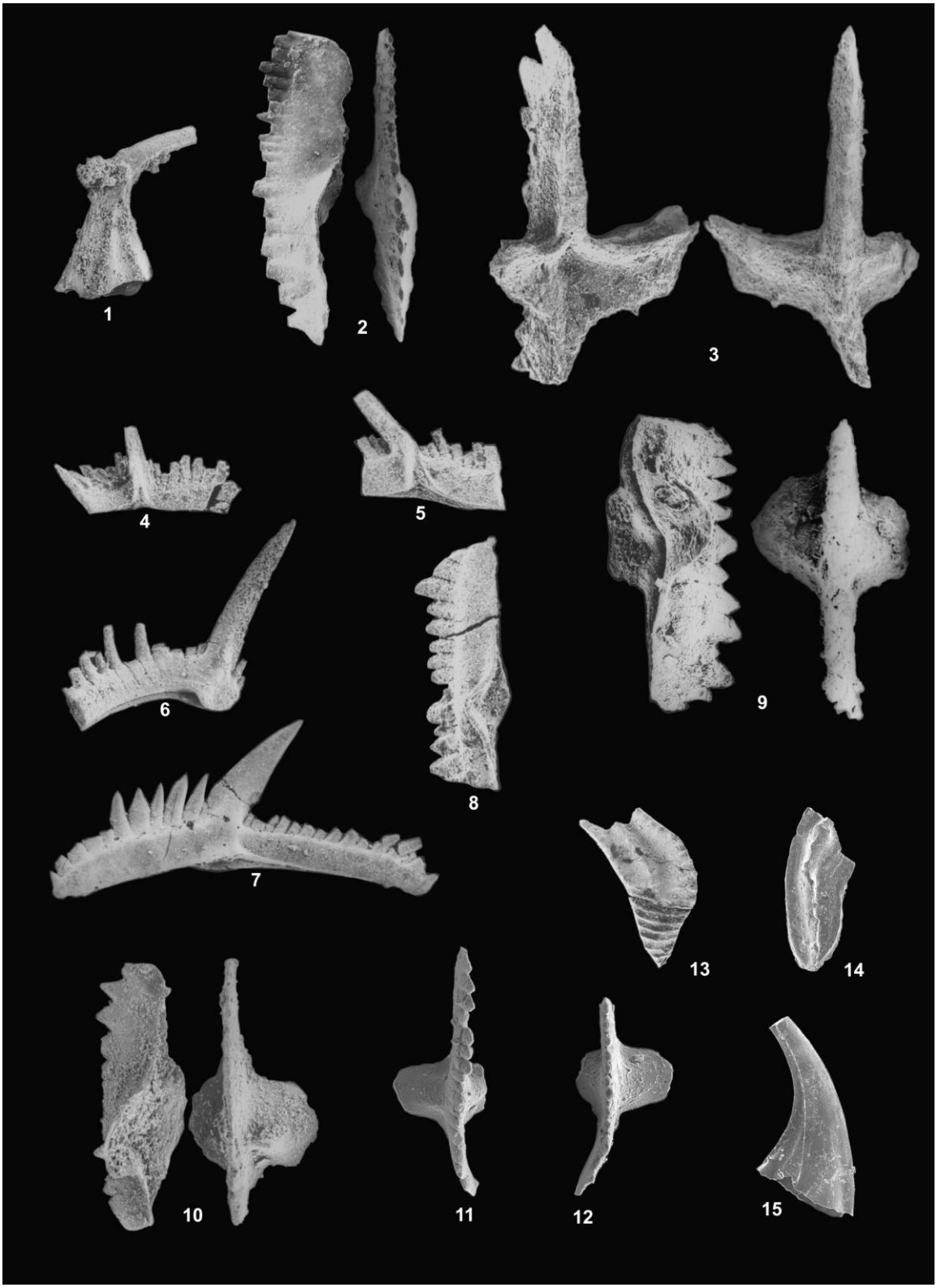
Paleozoic Succession of the Castillejos-Turón Unit

In the Alpine Castillejos-Turón Unit, the Morales and Santi Petri formations are lacking, as they are exclusively present in the Ardales-Mojón Unit (figs. 4, 5). The Paleozoic rocks of the Castillejos-Turón Unit are represented by small outcrops of pelites and conodont-bearing Silurian-Devonian nodular to very-thin-bedded limestones (Kockel 1959). These limestones (11, col. D, in fig. 5) are quite similar to those of the Ardales Castle Rock. Preliminary CAI data from some conodont elements isolated from these limestones show values around 5. These are followed by pelites and fine-grained sandstones (12 in fig. 5), by black cherts and limestones of the Carboniferous Falcoña Formation (6 and 7 in fig. 5), and by the Culm facies turbiditic pelites, graywackes, and conglomerates of the post-Viséan Almogía Formation (8 and 9 in fig. 5). The top of the Paleozoic succession is represented by the Upper Carboniferous Marbella Formation (10 in fig. 5), which deeply erodes the underlying deposits. It is mainly formed by matrix-supported, poorly sorted, coarse and massive conglomerates (debris flow deposits), including cobbles and boulders eroded from the underlying Paleozoic succession. Clast-supported conglomerates and finer-grained facies, such as coarse-grained gravelly sandstones, are also present but only subordinately. A distinctive feature of this formation is the common occurrence of big clasts (cobbles and boulders) of limestones containing Carboniferous (Viséan and Bashkirian) shallow-water fossils (Buchroithner et al. 1980; Herbig 1984) and of felsic plutonic and metamorphic rocks of unknown provenance.

Mesozoic-Cenozoic Succession of the Ardales-Mojón Unit

The Meso-Cenozoic cover of both Alpine Malaguide units starts with continental red pelites and quartz-rich sandstones with intercalated channelled conglomerates (13 and 14 in fig. 5), belonging to the Middle-Upper Triassic Saladilla Formation (Martín-Algarra et al. 1995). They show "Pseudo-Verrucano" lithofacies (Perrone et al. 2006; Critelli et al. 2008). Jurassic, Cretaceous, and Tertiary Malaguide rocks are also widely present in the outcrops around Ardales, usually detached from underlying deposits.

In the Ardales-Mojón Unit, the cover is mainly represented by Triassic rocks, which form a nearly continuously outcropping belt of an average stratigraphic thickness of around 150 m (fig. 4). The Saladilla Formation redbeds rest unconformably either on the Ermita or on the Peña del Castillo successions, sealing the contact between them just a few hundred meters to the southwest of Ardales, in the Arroyo de las Viñas area (fig. 6). The Jurassic succession belonging to the Ardales-Mojón Unit mainly occurs 4 km to the northeast of Ardales, in the Sierra Blanquilla area (fig. 4). The succession is tectonically detached from its Triassic substratum. It is formed by shallow-marine dolomites and limestones of Hettangian-Sinemurian age, followed by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red limestones with Pliensbachian ammonites (16 and 17, col. A, in fig. 5). Just west of Ardales (fig. 6), lower Eocene limestones (Maaté et al. 2000) rest directly on Jurassic limestones. These deposits are substantially thinned and found only in very small and scattered outcrops. They are followed by Aquitanian pelites rich in planktonic and benthic foraminifera that belong to the Pantano de Andrade Formation (Bourgois 1978). This succession also includes brown-yellowish immature sandstones and conglomerates with clasts, boulders, and tens-of-meters-sized blocks, consisting of Paleozoic pelites, graywackes, and conglomerates (González-Donoso et al. 1986; Martín-Algarra 1987). These rocks are also intensely imbricated with Triassic and Paleozoic deposits and are referable to the Upper Oligocene–upper Aquitanian, synorogenic Ciudad Granada Group of the Malaguide Complex (Guerrera et al. 1993; Martín-Algarra et al. 2000).

Mesozoic-Cenozoic Succession of the Castillejos-Turón Unit

The Mesozoic succession unconformably overlying the Paleozoic rocks of the Castillejos-Turón Unit also starts with Triassic continental red pelites, sandstones, and conglomerates of the Saladilla Formation, reaching up to 100 m in thickness (fig. 5). Gypsum is locally present in the upper part of the succession (14, col. D, in fig. 5). The Jurassic and Cretaceous succession of the Castillejos-Turón Unit is thicker and stratigraphically more complete than that of the underlying Ardales-Mojón Unit. Hettangian dolostones and Sinemurian platform limestones are followed by pelagic Upper Pliensbachian–Toarcian to Upper Jurassic cherty limestones and Lower Cretaceous marly limestones and marls sometimes bearing ammonites (18 and 19 in fig. 5; Kockel 1963; Martín-Algarra 1987). The succession ends with carbonate breccias, brownish-orange marly pelites, lithic-calcareous sandstones, turbiditic sandstones, and channelled conglomerates—including clasts of Meso-Cenozoic, Triassic, and Paleozoic rocks—belonging to the Aquitanian Pantano de Andrade Formation (21 and 22 in fig. 5). This formation widely crops out in the Loma del Castillo de Turón, along a narrow, NE-SW-oriented belt of subvertical to gently NW-dipping strata, also including hundreds-of-meters-sized blocks of Jurassic limestones and marls (fig. 4).

Discussion

The stratigraphical and paleontological data from the Malaguide rocks cropping out around Ardales confirm the Early Silurian–Devonian ages previously reported by Kockel (1959), Kockel and Stoppel (1962), and Rodríguez-Cañero (1993*a*). In addition, Upper Ordovician and Lower Carboniferous conodont associations are also reported in this study. The recognized ages necessarily imply the existence of at least one tectonic contact in the geological cross section observable along the southern slope of the Peña del Castillo, because Lower-Middle Devonian pelagic thin-bedded nodular limestones of the Ermita succession are covered by Lower Silurian to Lower Devonian limestones of the Peña del Castillo succession. Consequently,

Figure 8. Conodonts obtained from Malaguide rocks of the Ardales-Mojón Alpine Unit at and around of the Ardales Castle site. *1*, *Dapsilodus obliquicostatus* (Branson & Mehl 1933). Element S_1-S_2 707-183, sample 183, ×120. Base of the Peña del Castillo Subunit. *2*, *Ozarkodina sagitta sagitta* (Walliser 1964). Element P_1 710-183, sample 183, ×150. Peña del Castillo Subunit. *3*, *Kockelella manniki* Serpagli & Corradini 1998. Element P_1 706-184, sample 184, ×80. Peña del Castillo Subunit. *4*, *Ozarkodina excavata excavata* (Branson & Mehl 1933). Element S_1 1028-309, sample 309, ×70. Peña del Castillo Subunit. *5*, *Ozarkodina excavata excavata* (Branson & Mehl 1933). Element S_2 1029-309, sample 309, ×70. Peña del Castillo Subunit. *6*, *Ozarkodina excavata excavata* (Branson & Mehl 1933). Element M 1027-309, sample 309, ×60. Peña del Castillo Subunit. *7*, *Ozarkodina excavata excavata* (Branson & Mehl 1933). Element P_2 1026-309, sample 309, ×60. Peña del Castillo Subunit. *8*, *Ozarkodina excavata excavata* (Branson & Mehl 1933). Element P_1 1023-309, sample 309, ×60. Peña del Castillo Subunit. *9*, *Ozarkodina remscheidensis remscheidensis* (Ziegler 1960). Element P_1 1062-311, sample 311, ×80. Peña del Castillo Subunit. *10*, *Ozarkodina pandora* morphotype ε Murphy, Matti & Walliser 1981. Element P_1 1022-308, sample 308, ×100. Peña del Castillo Subunit. *11*, *Ozarkodina steinhornensis miae* (Bultynck 1971). Element P_1 88A-187-2, sample 187, ×100. Ermita Subunit. *12*, *Ozarkodina steinhornensis miae* (Bultynck 1971). Element P_1 88A-187-1, sample 187, ×110. Ermita Subunit. *13*, *Polygnathus linguiformis*? Hinde 1879. Element P_1 1010-186, sample 186, ×45. Ermita Subunit. *14*, *Amorphognathus ordovicicus*? Branson & Mehl 1933. Fragment of element P_1 03A-69-6, sample 03A-69, ×100. Ermita Subunit. *15*, *Scabbardella altipes* (Henningsmoen 1948). Element a 03A-69-10, sample 03A-69, ×100. Ermita Subunit.

both successions must be considered independent, stacked tectonic subunits within the Ardales-Mojón Alpine Unit.

The undated black cherts located on top of the Ermita succession (7 in fig. 7*B*), however, include an Upper Ordovician block close to its base. As these black cherts are sandwiched between Upper Ordovician and Lower Silurian horizons well dated by conodonts (6 and 8, respectively, in fig. 7*B*), they could be considered to be of latest Ordovician or earliest Silurian (Llandovery) age. In this hypothesis, the tectonic contact must be located on top of the Devonian limestones (3 in fig. 7*B*), just below the Ordovician block and immediately above the brecciated horizon (4 in fig. 7*B*) and then within the iron-rich horizon at the very base of the black cherts (5 in fig. 7*B*). Nevertheless, cherts of latest Ordovician or earliest Silurian age have never been found, up to now, either in the Betic Cordillera or in the Moroccan Rif. On the contrary, the only well-dated black cherts in the Betic-Rifian Malaguide-Ghomaride complexes are the Lower Carboniferous (Tournaisian) radiolarites of the Falcoña Formation (O'Dogherty et al. 2000). If the hypothesis of a Carboniferous age for the cherts below the Castle Rock is accepted, then the tectonic contact between the Ermita and the Peña del Castillo subunits must be located at the base of dated Silurian limestones just on top of the black cherts. Detailed structural analysis of this upper surface supports this latter interpretation (Martín-Algarra et al. 2009).

The presence of a conodont-bearing Viséan limestone horizon attached to cherts of the Ermita Subunit in the Arroyo de las Viñas section (6 and 7, col. C, in fig. 5), a few hundred meters to the southwest of Ardales (fig. 6), supports a Carboniferous age for the cherts outcropping just below the Castle Rock. Moreover, the succession of the Ermita Subunit along the Arroyo de las Viñas is quite similar to that described in the Falcoña-type section (Herbig 1983). There, some discontinuous lenses of basic volcanic rocks are also present within the pelites just below the cherts; the latter are overlain by Viséan limestone beds, well dated by conodonts (Rodríguez-Cañero and Guerra-Merchán 1996) that constitute the upper member of the Falcoña Formation. Consequently, the black chert succession at the base of the Peña del Castillo is well correlatable with the Falcoña Formation. The absence, at the top of the cherts, of both Viséan limestones and clastics of the Almogía Formation may be reasonably considered as being a result of tectonic elision.

Several observed features suggest a sedimentary origin for the inclusion of the Ordovician block (6 in fig. 7*B*) within the lower part of the cherts: (1) the block is not surrounded by any striated fault surface or fault gauge; (2) the iron-rich patinas and nodules present in the lowest chert beds (5 in fig. 7*B*) can be interpreted as being due to stratigraphic condensation and hardground formation; and (3) the presence of breccias surrounding the underlying Devonian tentaculites limestone (4 in fig. 7*B*), the sharp contact on top of the latter (3 in fig. 7*B*), as well as its wedge-shaped morphology and internal deformation by stretching and shearing, can partially be related to submarine sliding and resedimentation before deposition of cherts and later tectonic overprint. A stratigraphic condensation related to the Late Devonian Kellwasser Event is well documented in other sites of the Malaguide Complex (Rodríguez-Cañero 1993*b*, p. 14; 1995, p. 8). In addition, at that time, frequent resedimentation processes have been recognized in Paleozoic successions of Minorca (Balearic Islands) that show great resemblances to Malaguide successions (Herbig 1985; Henningsen and Herbig 1990).

Therefore, the lower Alpine Malaguide Unit of the Ardales area (Ardales-Mojón Unit) includes two superposed tectonic subunits made of different Paleozoic successions. These successions actually characterize a lower Ermita Subunit, consisting of Lower-Middle Devonian sandstones, slates, and limestones, followed by probably Lower Carboniferous black cherts and an upper Peña del Castillo Subunit, made of mainly Silurian–Lower Devonian limestones overlaid by a few meters of undifferentiated Devonian-Carboniferous pelites and fine-grained sandstones. The contact between both subunits is sealed, west of Ardales (col. C in fig. 5; fig. 6), by Triassic redbeds of the Saladilla Formation. Consequently, the tectonic emplacement of the Peña del Castillo Subunit on the Ermita Subunit accounted in pre-Triassic time and must be considered Variscan. This is in contrast to the previous hypothesis of a sedimentary emplacement of Paleozoic limestones as a result of gravitational mass-flow processes, related to the Carboniferous deposits of the Marbella Formation, which are completely lacking close to Ardales. The Marbella Formation outcrop located south of Ardales, at the confluence of the Arroyo de la Torre with the Arroyo de las Moratillas, is tectonically sandwiched in a fault zone between the backthrusted Campo de Gibraltar Complex and older deposits of the Ardales-Mojón Unit (Ermita Subunit) and better interpreted as a tectonic slice belonging to the overlying Alpine Castillejos-Turón Unit. Actually, in this latter unit the Marbella Formation is very well developed along the Turón River valley, where

limestone cobbles and boulders are quite abundant. But such big clasts of limestone are made of Viséan to Bashkirian shallow-marine rocks (Buchroithner et al. 1980; Herbig 1984) of unknown provenance, whereas the existence of big clasts of Silurian-Devonian pelagic limestones, like those of the Peña del Castillo Subunit, has never been noticed within the well-studied clasts of limestone included within the Marbella Formation outcropping either to the west or to the south of Ardales.

Further information on the Paleozoic tectonometamorphic evolution is provided by CAI of studied conodonts. Most conodont elements of the Ardales-Mojón Unit (either from the Variscan lower Ermita Subunit or from the higher Peña del Castillo Subunit) show CAI around 5.5, except for conodonts obtained from the Ordovician block, which have lower CAI values (4–4.5). CAI values around 5 have also been observed in conodont elements obtained from Silurian-Devonian limestones of the Castillejos-Turón Alpine unit. Similar CAI values are quite frequently observed in conodont elements found in Malaguide outcrops other than Ardales (Rodríguez-Cañero 1993*a*; Navas-Parejo et al. 2008), as well as in equivalent Ghomaride units of the Rif (Rodríguez-Cañero et al. 1990; Martín-Algarra and Rodríguez-Cañero, unpublished results). Nevertheless, lower (4) and higher (6) CAI values have also been found in the Malaguide Complex (e.g., Rodríguez-Cañero 1995; Rodríguez-Cañero and Guerra-Merchán 1996; O'Dogherty et al. 2000). According to the most accepted models of CAI significance (Rejebian et al. 1987; Sarmiento and García-López 1996), the CAI values obtained in this study indicate that the Malaguide Paleozoic successions outcropping at Ardales have been affected by an anchizonal to epizonal metamorphic overprint.

A metamorphic overprint, similar to or slightly more intense than that indicated by conodont CAI values at Ardales, has been recognized in terrigenous-clastic Malaguide Paleozoic successions of the Montes de Málaga area. There, rocks belonging mainly to the Morales Formation (subordinate to the Santi Petri or higher formations) provide very-low-grade to low-grade and occasionally medium-grade metamorphic mineral associations, indicating temperatures (T) up to 300°–500°C and pressure up to 5 kbar (Ruiz-Cruz and Andreo 1996*b*; Ruiz-Cruz 1999, 2001; Ruiz-Cruz and Galán 2002; Ruiz-Cruz and Nieto 2002; Ruiz-Cruz and Novák 2003). Mineral associations from Triassic Malaguide sediments in the same area indicate deep diagenesis to anchimetamorphic conditions, with $T > 140$°C and in some instances $T > 300$°C, with maximum pressure around 2.7 ± 3 kbar (Ruiz-Cruz and Andreo 1996*a*; Ruiz-Cruz 2004). This abrupt decrease in temperature from basement to cover is related to contrasting intensity and type of deformation, equivalent to that observable at Ardales, between Paleozoic and Meso-Cenozoic successions. Finally, in the studied area northeast of Ardales, Platt et al. (2003) have obtained one late Paleozoic zircon fission-track age (280 ± 19 Ma) and two latest Oligocene–early Miocene apatite fission-track ages (25.0 ± 4.0 and 22.0 ± 4 Ma) from samples taken at the base of the Morales Formation, near the contact with the underlying Alpujarride Complex. In brief, CAI data obtained in the Paleozoic successions of the Ardales area confirm that the Paleozoic successions of the Malaguide Complex were affected by pre-Alpine metamorphism, whereas both Paleozoic and Meso-Cenozoic cover record a complete Alpine diagenetic-metamorphic zonation (Ruiz-Cruz 2004).

Concluding Remarks

In the Ardales area, two Alpine Malaguide units can be recognized and mapped: the Ardales-Mojón Unit, in the footwall, and the Castillejos-Turón Unit, in the hanging wall. Both units consist of Paleozoic rocks unconformably covered by Triassic to upper Aquitanian sedimentary successions.

The Castillejos-Turón Unit consists of Silurian to Carboniferous rocks, represented mainly by Culm facies, post-Viséan turbiditic pelites, graywackes, and conglomerates (Almogía and Marbella formations) and by minor Lower Carboniferous cherts and limestones (Falcoña Formation), Devonian tentaculitid limestones, and Silurian nodular limestones. The Meso-Cenozoic cover of this Alpine unit is thicker and stratigraphically more complete than that of the underlying Ardales-Mojón Unit (fig. 5). In addition, the Jurassic to Aquitanian part of the succession is strongly tectonized and sheared along the Internal-External Domain boundary and frequently detached from its Paleozoic to Triassic substratum.

The new paleontological data reported in this article allowed us to detail the Paleozoic stratigraphy of the lower Alpine Malaguide Unit (Ardales-Mojón Unit). This unit includes two different Paleozoic successions that belong to two Variscan tectonic subunits. The Peña del Castillo Subunit is made of Silurian–Lower Devonian limestones, occasionally covered by undifferentiated Devonian-Carboniferous pelites and fine-grained sandstones. It overlies the Ermita Subunit, consisting of Lower-Middle Devonian (and older?) sandstones, pelites, and lime-

stones and Lower Carboniferous black cherts (including one block of Ordovician limestone).

The tectonic contact between these subunits is sealed by Triassic continental redbeds. Therefore, the thrusting of the Peña del Castillo Subunit on the Ermita Subunit occurred before the Triassic and must be considered Variscan. This tectonic superposition of two different Paleozoic successions may suggest a general reevaluation of the role of pre-Alpine structures within the Internal Domains of the Betic Cordillera. Besides, the CAI data reveal significant low-grade pre-Alpine metamorphism in the Malaguide Paleozoic rocks, in addition to the widely accepted existence of a weak alpine metamorphic overprint.

The new stratigraphic data of this article provide useful information for understanding the paleogeographical evolution of the Malaguide Paleozoic successions. The Malaguide Complex represents the southernmost wide piece of pre-Mesozoic sedimentary strata of the Iberian Peninsula, and Iberian Paleozoic terranes are a keystone for understanding the formation of the Variscan orogenic collage, the amalgamation of continental crust terranes that produced Pangea, and the closure of the two large, long-lived Paleozoic oceans, precursory of central and southern Western European and North African regions: the Rheic Ocean and the Paleotethys. The closure of the Rheic Ocean, responsible for the Variscan orogenesis in most Iberian Paleozoic terranes, accounted during Devonian–earliest Carboniferous times (Simancas et al. 2001, 2005; Arenas et al. 2007). The Paleotethys Ocean, however, closed much later, since Carboniferous time up to the end of the Triassic (Stampfli and Borel 2004).

Variscan tectogenesis in the Malaguide-Ghomaride Complex did not start before the latest Devonian and did not end before Late Carboniferous (post-Bashkirian) time. In addition, Malaguide successions are stratigraphically quite different from other coeval Ordovician to Carboniferous successions in southern Spain (for a recent updated and synthetic revision, see different contributions in Vera 2004). These dissimilarities support previous classic ideas on paleogeographical relations of the Internal Domains of the Betic Cordillera with regions located far to the east, such as the Alps, rather than with other Iberian regions. It is noteworthy that the Paleozoic successions of the Alps were related to the Paleotethys (von Raumer et al. 2002) whereas most Iberian and northwestern African Paleozoic terranes were related to the Rheic Ocean. In brief, the Paleozoic successions of the Malaguide Complex are not paleogeographically related to the Rheic Ocean but to the western prolongation of the Paleotethys (O'Dogherty et al. 2000), as part of the so-called Hun "superterrane" (von Raumer et al. 2002, 2003). The final welding of Gondwana and Laurasia was related to the closing of the western Paleotethys since mid-Carboniferous times (Stampfli and Borel 2004) and to deformation of the Paleozoic successions that are found today in the Alps (von Raumer 1998) and in the Internal Domains of the western Mediterranean Alpine Belts. A few millions of years after this continental collision, late Variscan extensional tectonics was associated with widespread Late Carboniferous–Permian magmatism and, finally, with the beginning of the Pangea fragmentation.

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First Late Ordovician conodont fauna in the Betic Cordillera (South Spain): a palaeobiogeographical contribution

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ABSTRACT

The youngest Ordovician conodont fauna in SW Europe has been found in the Malaguide Complex of the Betic Cordillera, SE Spain. It is also the first Ordovician conodont fauna in the Western Mediterranean Alpine Orogen. The conodont association, attributed to the Hirnantian (upper part of the *Amorphognathus ordovicicus* Biozone), is characterized by the predominance of *Walliserodus amplissimus* and *Scabbardella altipes* and by the absence of *Sagittodontina* and *Istorinus*, typical of the Mediterranean Province. This fauna differs markedly from those of the same biozone recorded in the Spanish Variscan Orogen of the Iberian Massif, which are attributed to the Katian. The Malaguide fauna shows, however, striking similarity to faunas of the Carnic Alps and some resemblance to those of the Pyrenees, Northern England and North Wales. These features suggest that Palaeozoic terranes of the Betic Cordillera were located far to the east of their present location and displaced westward during the Alpine Orogeny.

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Introduction

In southernmost Spain, Palaeozoic rocks, usually metamorphic, are exposed only in the Internal Domain of the Betic Cordillera that is the westernmost European Alpine Belt (Herbig, 1984; Martín-Algarra, 1987; Martín-Algarra *et al*., 2004). Nonetheless, thick terrigenous-clastic sedimentary Palaeozoic successions are present in the Malaguide Complex, where Llandovery to Bashkirian deposits have been dated mainly by conodonts. Conodonts are, consequently, essential tools for improving the poorly known Palaeozoic stratigraphy, and understanding the very complicated Alpine and pre-Alpine tectonic structure and geological history of this belt.

Silurian, Devonian and Carboniferous conodonts were reported by Kockel (1959) and Kockel and Stoppel (1962) in the western outcrops of the Malaguide Complex and, a few years later, in the eastern regions (van den Boogaard, 1965). Herbig (1984, 1985) identified several Middle and Upper Devonian and Carboniferous biozones; Rodríguez-Cañero (1993a) demonstrated the presence of many conodont biozones of the Middle and Upper Silurian, the Lower, Middle and Upper Devonian, and the Lower Carboniferous (Table 1). Most of these conodonts were obtained from: (i) tectonically isolated and scattered Silurian and Devonian carbonate outcrops with stratigraphic successions that vary from a few metres to several tens of metres in thickness (Kockel, 1959; Kockel and Stoppel, 1962; Herbig, 1984, 1985; Rodríguez-Cañero, 1993a,b); (ii) Lower Carboniferous thin siliceous and carbonate deposits intercalated between unfossiliferous turbiditic slates and greywackes (Herbig, 1983; Rodríguez-Cañero and Guerra-Merchán, 1996; O'Dogherty *et al*., 2000; Navas-Parejo *et al*., 2008); and (iii) centimetre- to few metre-sized Devonian–Carboniferous carbonate clasts included within Carboniferous conglomerates (Buchroithner *et al*., 1980; Herbig, 1984).

The existence of Ordovician deposits in the Malaguide Complex of the Betic Cordillera was suspected (Martín-Algarra *et al*., 2004) but, until now, not biostratigraphically demonstrated. The present study reveals the existence of Upper Ordovician strata, which have been dated by conodonts of the *Amorphognathus ordovicicus* Biozone for the first time in the Betic Cordillera. In spite of its local character, this discovery is relevant because it represents the first solid evidence of the presence of Ordovician rocks in the Palaeozoic successions of the Western Mediterranean Alpine Chains that extend from southern Italy to Gibraltar. Consequently, this allows correlation with the well-known succession in the Alps and in other North-Gondwanic areas, and with the new standard chronostratigraphic scale (Bergström *et al*., 2009).

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Geological setting

As is usual for mountain belts in the Western Mediterranean Alpine Orogenic System, the Betic Cordillera is divided into tectonic units belonging to the External, Flysch and Internal Domains (Durand-Delga and Fontboté, 1980; Vera and Martín-Algarra, 2004). As a consequence of the post-Palaeozoic Pangaea breakoff, these domains constituted, from mid-Jurassic time, the northern (External) and southern (Internal) continental margins of a narrow, deep marine (Flysch) basin that connected the western Tethyan and central Atlantic oceans. Cretaceous to Miocene subduction and metamorphism followed by continental collision transformed these former margins and basin, and their basements, in the Alpine stack of thrust nappes that constitute the Betic Cordillera today.

Palaeozoic terranes are practically absent in the External and Flysch

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Table 1 Conodont biozones in the Malaguide Complex after Rodríguez-Cañero (1993a). Numbers correspond to samples.

Epoch/period	Stage/Epoch	Localities / Conodont biozones	Arroyo de las Viñas	Peña del Castillo de Ardales	Almogía I	Almogía II	Almogía IIIA	Almogía IIIB	Almogía IV	Fuengirola	Arroyo de Casa Otero
Lower Carboniferous	Visean	*bilineatus*									103
		transatlanticus									
		homopunctatus									
	Tournaisian	*anchoralis-latus*									
		typicus									
		isosticha-crenulata									
		sandbergi									
		duplicata									
		sulcata									
Upper Devonian	Famennian	*praesulcata*									
		expansa									
		postera									
		trachytera								167	
		marginifera								165	
		rhomboidea							230		
		crepida u.m.	190					102	231		
		crepida u.						89			
		crepida m.						87	232		
		crepida l.							238		
		triangularis u.							233		
		triangularis m.					305				
		triangularis l.					94				
	Frasnian	*linguiformis*					92		229		
		rhenana									
		jamiae									
		hassi									
		punctata									
		transitans			70					172	
		falsiovalis									
Middle Devonian	Givetian	*disparilis*			74						
		hermanni									
		varcus u.									
		varcus m.			78						
		varcus l.				86					
		ensensis				110					
	Eifelian	*kockelianus*				82					
		costatus									
		partitus									
Lower Devonian	Emsian	*patulus*									
		serotinus									
		inversus-laticostatus									
		gronbergi	200	187							
		dehiscens									
		kitabicus									
	Pragian	*pirenae*									
		kindlei									
		sulcatus		308							
	Lockovian	*pesavis*									
		delta	197								
		eurekaensis	196	311							
		postwosch/woschmidti									
Silurian	Pridoli	*eosteinhornensis*									
	Ludlow	*crispa*		185							
		latialata									
		siluricus		184							
		Ploeckensis									
	Wenlock	*crassa*									
		sagitta		183							
		patula									
		amsdeni-ranuliformis		182							
		amorphognathoides	189								
	Llandovery	*celloni*									
Ordovi.	Hirnantian	*ordovicicus*	69*								
	Katian										

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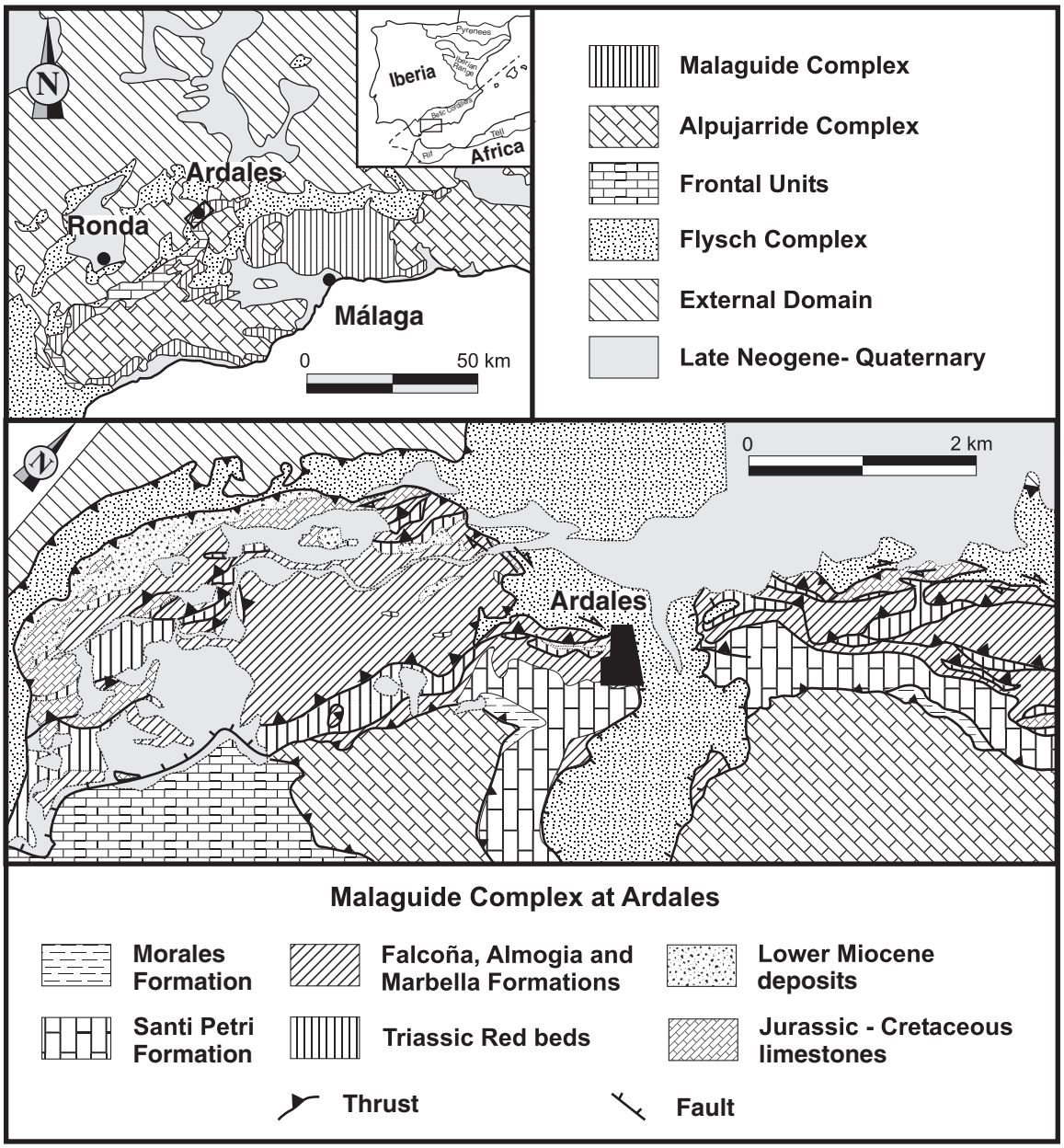

Fig. 1 Tectonic sketch of the Western Betic Cordillera and simplified geological map of the Malaguide Complex at Ardales.

Domains, as well as in the Frontal Units of the Internal Domain. They are, however, widely present in the other three tectonic complexes distinguished within the Internal Domain, which are named, from bottom to top, the Nevado-Filabride, Alpujarride and Malaguide complexes respectively. The Nevado-Filabride and Alpujarride Palaeozoic terranes consist exclusively of metamorphic rocks; pre-Mesozoic fossils have rarely been found there (Lafuste and Pavillon, 1976; Gómez-Pugnaire *et al.*, 1982). However, the Malaguide Palaeozoic terrain consists of thick, mainly siliciclastic and deep marine (turbiditic) successions bearing fossils (Martín-Algarra *et al.*, 2004, and references therein).

The Malaguide Complex is exposed widely at Ardales, 30 km to the NE of Málaga, where it forms a narrow, discontinuous and strongly tectonized klippen belt (Fig. 1). The Malaguide succession cropping out there (Fig. 2) consists of Ordovician to Upper Carboniferous rocks that were strongly deformed during pre-Alpine and Alpine events (Martín-Algarra *et al.*, 2009a). Palaeozoic rocks include turbidites and slates, with subordinate conglomerate, chert and pelagic limestone bearing conodonts, and make up part of two superimposed Variscan tectonic units with different Palaeozoic successions, assigned to the Ermita (lower) and Peña del Castillo (upper) units (Martín-Algarra *et al.*, 2009b).

From bottom to top, the studied succession in the Ermita Unit consists of (Figs 2 and 3): (a) a few metres of

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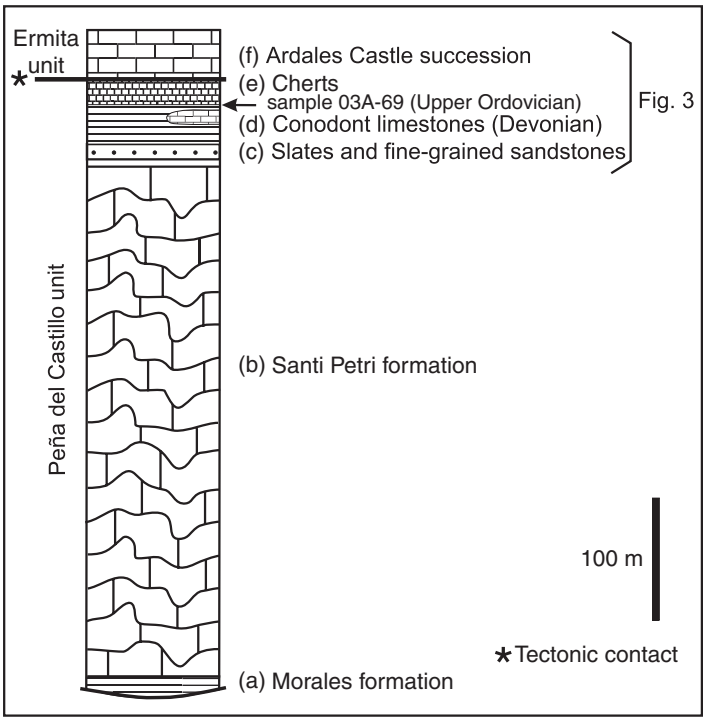

Fig. 2 Stratigraphy of the Malaguide Palaeozoic succession at Ardales.

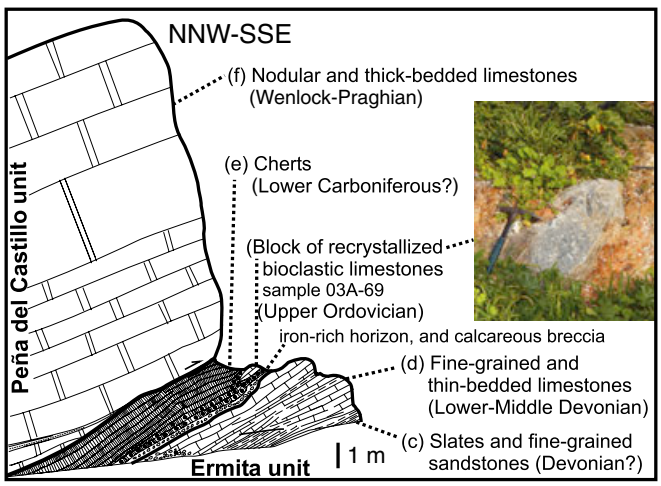

Fig. 3 The Ardales Castle Rock section and the location of the limestone conodont sample.

slightly metamorphic pre-Devonian slate and metasandstone belonging to the Morales Formation; (b) a thick and strongly folded succession of Devonian (?) calcareous slate and fine-grained turbiditic limestone of the Santi-Petri Formation; (c) a few tens of metres of slate and fine-grained sandstone with some discontinuous bodies of basic volcanic rock; (d) a few-metre-thick lens of grey-bluish fine-grained and thinly bedded pelagic limestone, with Lower to Middle Devonian conodonts and dacryoconarids (Rodríguez-Cañero, 1993a), which is bound by a thin brecciated horizon; and (e) a few metres of strongly sheared black radiolarian cherts, which are locally associated with bluish limestone bearing Lower Carboniferous (Visean) conodonts. This chert and limestone belonging to the Falcoña Formation is truncated by a pre-Alpine thrust fault that divides the top of the lower (Ermita) Variscan tectonic unit from the base of the upper (Peña del Castillo) Variscan unit.

Table 2 Late Ordovician conodont taxa from Ardales.

The Ardales castle rock section	
Sample number	03A – 69
Amount of dissolved sample	1975 g
Amorphognathus cf. *ordovicicus*	
Pa element	6 frag.
Pb element	3
Scabbardela altipes	
a element	9
b? element	7
e element	10
Scabbardella sp.	5
Walliserodus amplissimus	
a element	3
b element	2
c element	2
d element	2
e element	9
Fragments	18
Walliserodus cf. *amplissimus*	
d ? element	4
e element	1
Fragments	7
Walliserodus? sp.	7
Protopanderodus? sp.	1
Hamarodus europaeus	
Sc element	1
Eocarniodus	2
Gen et sp. indet.	9
Fragmentary specimens	38
Total	146

The Peña del Castillo Unit (Figs 2 and 3) is composed of Silurian–Lower Devonian nodular and thickly bedded limestone with conodonts (f), constituting the approximately 30-m-thick Ardales Castle succession (Rodríguez-Cañero, 1993a; Rodríguez-Cañero *et al.*, 1997).

Results and discussion

The Ordovician conodont association was found in a block of recrystallized bioclastic limestone that lies within a siliceous–ferruginous–calcareous claystone horizon (< 1 m thick) with discontinuous calcareous breccia, in the basal chert beds on top of the Middle Devonian pelagic limestone (levels e and d, above mentioned, respectively). The inclusion of the Ordovician block within surrounding sediments was probably the result of submarine sliding and resedimenta-

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tion (see discussion in Martín-Algarra *et al*., 2009a; p. 256). The sampled limestone block has a volume of about 0.3 m^3 (Fig. 3).

In total, 146 conodont elements were recovered; the recognized taxa are summarized in Table 2. Specimen abundance is moderately high, with 75 elements per kg of dissolved sample. Conodont element size ranges from 0.2 to 1 mm, thus indicating that hydrodynamic sorting was minimal. The specimens are well preserved in general: only a few are fragmentary; the colour alteration index of Epstein *et al*. (1977) is 4–4.5.

The low-diversity conodont fauna from Ardales includes *Walliserodus* Serpagli, 1967; *Scabbardella* Orchard, 1980; *Amorphognathus* Branson and Mehl, 1933; *Hamarodus* Viira, 1974 and *Protopanderodus*? Lindström, 1971 (Figs 4 and 5). Coniform elements comprise more than 80% of the conodont collection; *Walliserodus*, including forms closely related to Silurian species, dominates the fauna (Figs 4 and 5). The genera *Sagittodontina* Knüpfer, 1967 and *Istorinus* Knüpfer, 1967 are, however, absent, although they are typical genera of the Mediterranean Province and indicative of the polar to subpolar regions of higher latitudes around Northern Gondwana (Sweet and Bergström, 1984). Consequently, the Late Ordovician fauna found at Ardales cannot be ascribed to any of the Ordovician palaeobiogeographic provinces defined by Bergström (1973, 1990) and Sweet and Bergström (1984), neither can it be ascribed to any of the biofacies proposed by Sweet and Bergström (1984).

The conodont association of the Malaguide Complex at Ardales is very peculiar and differs significantly in taxonomic composition from associations in other areas of the Variscan Orogen of the Iberian Peninsula dated as Katian, where *Sagittodontina* Knüpfer, 1967 and *Istorinus* Knüpfer, 1967 are common (Fuganti and Serpagli, 1968 Carls, 1975; Hafenrichter, 1979; Sarmiento, 1990, 1993; Sarmiento *et al*., 1995, 1999, 2001, 2008; Sarmiento and Gutiérrez-Marco, 1999; Del Moral, 2003, 2005, 2007; Del Moral and Sarmiento, 2008). These Katian conodonts have been recorded from several lithostratigraphic units belonging to different structural and palaeogeographic zones of the Variscan Iberian Massif (Table 3, Fig. 6): (i) the Cantabrian Zone (an unnamed formation at Portilla de Luna: Del Moral, 2003); (ii) the West Asturian Leonese Zone (La Aquiana Fm.: Sarmiento, 1993); (iii) the Iberian Cordillera (Cystoid Limestone and Ojos Negros formations: Carls, 1975; Hafenrichter, 1979; Del Moral, 2007); (iv) the Pyrenees (El Baell Fm.: Sanz-López and Sarmiento, 1995; and references therein); (v) the Catalonian Coastal Ranges (Madremanya Limestone: Sarmiento *et al*., 1995); (vi) the Central Iberian Zone (Casaio Fm., limestone pebbles of the Rozadais Fm., Urbana Limestone, Ferradosa Fm., and Adamuz Variscan olistoliths: Fuganti and Serpagli, 1968; Hafenrichter, 1979; Sarmiento, 1990, 1993; Sarmiento and Gutiérrez-Marco, 1999; Sarmiento *et al*., 2001; Del Moral, 2007; Del Moral and Sarmiento, 2008); and (vii) the Ossa-Morena Zone (Pelmatozoan Limestone: Hafenrichter, 1979; Sarmiento *et al*., 2008).

The Ardales conodont fauna also differs from those of many other Southern European and North African sites: Thuringia (Knüpfer, 1967; Ferretti and Barnes, 1997), NW France (Lindström and Pelhate, 1971; Weyant *et al*., 1977), Sardinia (Ferretti and Serpagli, 1999) and Libya (Bergström and Massa, 1992), where *Sagittodontina* and *Istorinus* are common but *Walliserodus* is scarce or absent.

Nevertheless, Late Ordovician conodont faunas from the Carnic Alps (Schönlaub, 1988; Bagnoli *et al*., 1998; Ferretti and Schönlaub, 2001), Northern England and North Wales (Orchard, 1980), and the Spanish Pyrenees (Sanz-López and Sarmiento, 1995) are similar to the Ardales fauna. Bagnoli *et al*. (1998) reported a more diverse conodont fauna without *Sagittodontina* and *Istorinus* from the Uggwa Formation of the Valbertad section (Carnic Alps). Ferretti and Schönlaub (2001) also noticed the absence of *Sagittodontina* and *Istorinus* and stated that (p. 9) *Walliserodus amplissimus becomes dominant in the uppermost horizons* of the Uggwa and Wolayer limestones of the Central Carnic Alps (Austria) and that (p. 10) *Important levels with Hirnantia fauna were reported at the top of the Uggwa Limestone, immediately before the beginning of the Plöcken Unit*. Orchard (1980) showed that *A. ordovicicus*?, *W. aff. amplissimus* and *Hamarodus europaeus* are the only conodonts present in Hirnantian levels of the Ashgill Series of Northern England and North Wales. Finally, in the associations from the Spanish Pyrenees, *Sagittodontina* and *Istorinus* are absent but *Walliserodus* is relatively abundant (Sanz-López and Sarmiento, 1995) (Table 3).

As previously indicated, the Ardales fauna contains *Walliserodus* elements with a morphology similar to that of Silurian species (Figs 4 and 5). This suggests a replacement of Ordovician taxa by Silurian-like faunas in the latest Ashgill, as proposed by Barnes and Bergström (1988).

The Ardales conodonts correspond to generalist and eurythermal species from the Atlantic Faunal Region, as

Fig. 4 Late Ordovician conodonts of the Malaguide Complex at Ardales 1. *Walliserodus amplissimus* (Serpagli, 1967), a element; 03A-69-58. (a) Inner lateral view; (b) outer lateral view. 2. *Walliserodus amplissimus* (Serpagli, 1967, b element; 03A-69-60. (a) Inner lateral view; (b) outer lateral view. 3. *Walliserodus amplissimus* (Serpagli, 1967), c element; 03A-69-50. (a) Inner lateral view; (b) outer lateral view. 4. *Walliserodus amplissimus* (Serpagli, 1967), d element; 03A-69-40. (a) Outer lateral view; (b) inner lateral view. 5. *Walliserodus amplissimus* (Serpagli, 1967), e element; 03A-69-57. (a) Inner lateral view; (b) outer lateral view. 6. *Walliserodus amplissimus* (Serpagli, 1967), e element; 03A-69-5. (a) Inner lateral view; (b) outer lateral view. 7. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), a element; 03A-69-29. (a) Outer lateral view; (b) inner lateral view. 8. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), b element; 03A-69-39. (a) Inner lateral view; (b) outer lateral view. 9. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), a element; 03A-69-26. (a) Inner lateral view; (b) outer lateral view. 10. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), a element; 03A-69-17. (a) Inner lateral view; (b) outer lateral view. 11. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), b element; 03A-69-59. (a) Inner lateral view; (b) outer lateral view. 12. *Walliserodus* cf. *amplissimus* (Serpagli, 1967), e element; 03A-69-2. (a) Outer lateral view; (b) inner lateral view. White drawings correspond to the basal sections. Scale bars = 200 μm.

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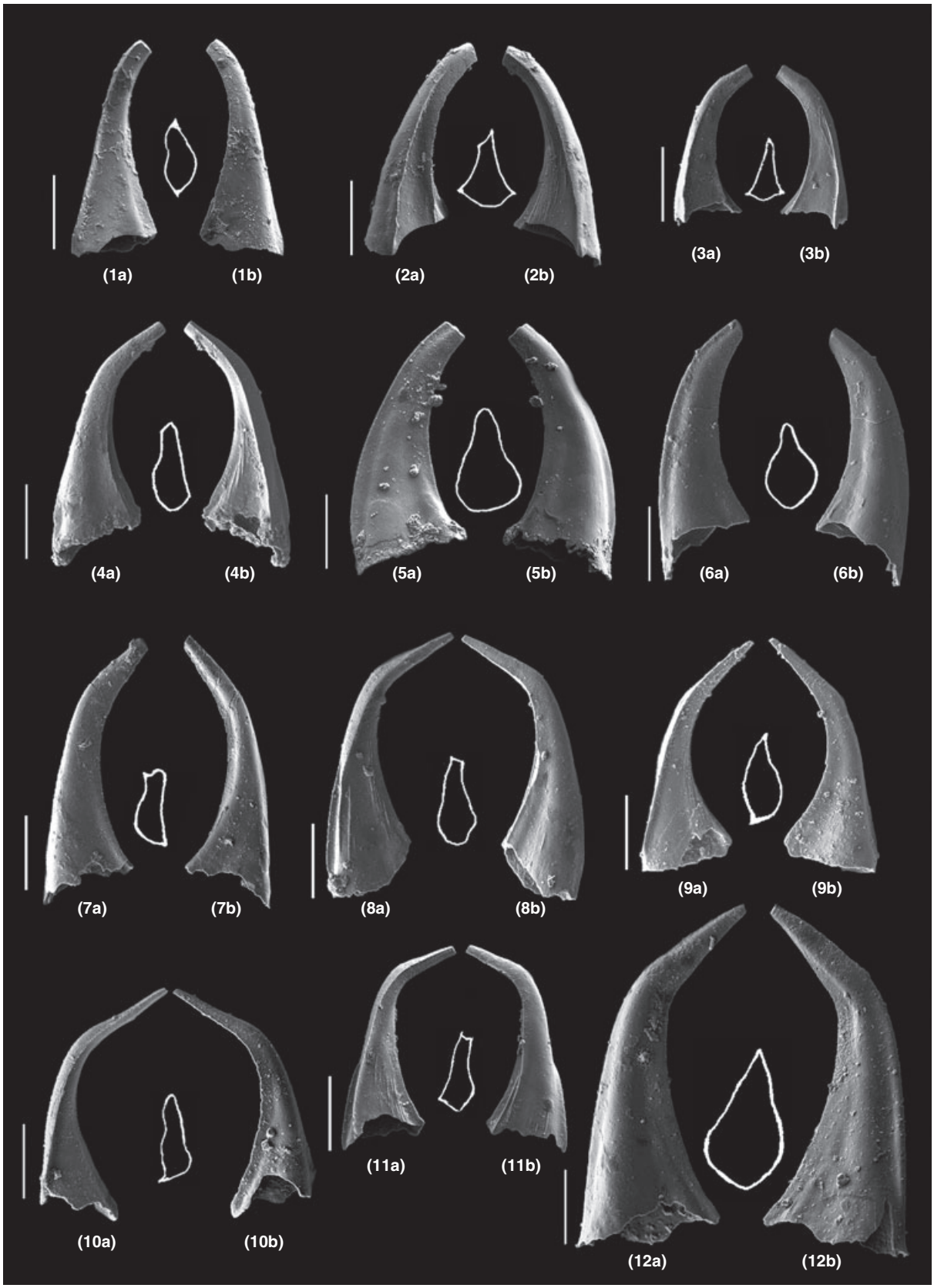

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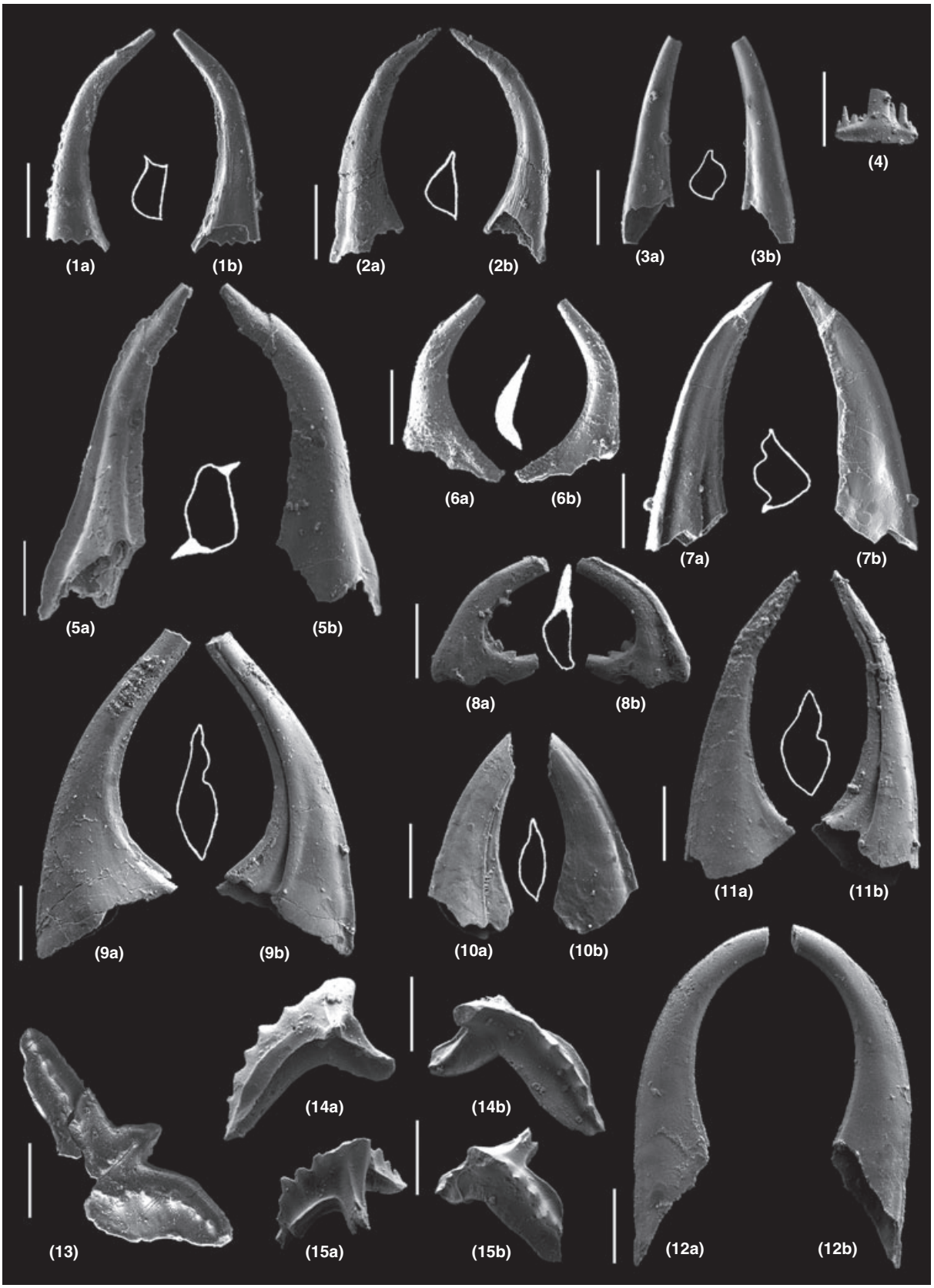

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indicated by their low diversity and by the predominance of coniform taxa (Bergström, 1990). This suggests that the Malaguide fauna evolved under ecologically stressed conditions, and that only species adapted to rapidly changing environments survived. Hirnantian conodont associations from the Atlantic Faunal Region are, in general, very scarce and poorly diversified (Orchard, 1980; Bergström and Orchard, 1985; Barnes and Williams, 1988). Moreover, they are absent from Hirnantian strata in other Spanish sites (Iberian Massif), where the sediments are dominated by glaciomarine diamictites (Carls, 1975; Robardet, 1981; Fortuin, 1984; and Robardet and Doré, 1988, among others).

Table 3 Late Ordovician conodont associations in the Iberian Peninsula, with the total number of elements so far published for each area, to enhance the differences in abundance of some genera (e.g. *Walliserodus*) between the (Alpine) Malaguide Complex of the Betic Cordillera (*this paper, Hirnantian) and other (Variscan) Iberian sites (Katian).

Late Ordovician conodont associations in Iberian Peninsula	Cantabrian Zone	West Asturian Leonese Zone	Iberian Cordillera	Central Iberian Zone					Ossa-Morena Zone	Catalonian Coastal Ranges		Betic Cordillera
	Unnamed Fm. Portilla de Luna	La Aquiana Fm.	Cystoid Limestone and Ojos Negros Fms.	Casaio Fm.	Limestone pebbles of the Rozadais Fm.	Urbana Limestone Fm.	Ferradosa Fm. (Portugal)	Adamuz (Córdoba)	Pelmatozoan Limestone	Madremanya Limestone	El Baell Fm. Pyrenees	Malaguide Complex*
Sagittodontina		3	326	1	134	1413	9		53			
Istorinus			56			106	1	2	3			
Eocarniodus			36			2057	4		2	1		1
Icriodella	1		14			74		4	1	2		
Amorphognathus	7	1	418		24	1933	4		37	6	8	9
Hamarodus			19	1	11	79			2			
Protopanderodus												1
Panderodus	2		221			176	3			1	7	1
Scabbardella	3	2	515	6	17	3229	6	1	101	3	4	31
Walliserodus						2	1	6	1		3	55
Gen et sp. indet.								1				9

Fig. 5 Late Ordovician conodonts of the Malaguide Complex at Ardales 1. *Walliserodus*? sp., 03A-69-7. (a) Outer lateral view; (b) inner lateral view. 2. *Walliserodus*? sp., 03A-69-8. (a) Outer lateral view; (b) inner lateral view. 3. *Walliserodus*? sp., 03A-69-12. (a) Outer lateral view; (b) inner lateral view. 4. Eocarniodus element, 03A-69-104. 5. Gen et sp. indet., 03A-69-61. (a) Inner lateral view; (b) outer lateral view. 6. *Protopanderodus*? sp., 03A-69-16. (a) Outer lateral view; (b) inner lateral view. 7. Gen et sp. indet., 03A-69-36. (a) Inner lateral view; (b) outer lateral view. 8. *Hamarodus europaeus* (Serpagli, 1967), Sc element; 03A-69-106. (a) Inner lateral view; (b) outer lateral view. 9. *Scabbardella altipes* (Henningsmoen, 1948), a element; 03A-69-74. (a) Outer lateral view; (b) inner lateral view. 10. *Scabbardella altipes* (Henningsmoen, 1948), b? element; 03A-69-79. (a) Inner lateral view; (b) outer lateral view. 11. *Scabbardella altipes* (Henningsmoen, 1948), a element; 03A-69-74. (a) Outer lateral view; (b) inner lateral view. 12. *Scabbardella altipes* (Henningsmoen, 1948), e element; 03A-69-64. (a) Outer lateral view; (b) inner lateral view. 13. *Amorphognathus* cf. *ordovicicus* Branson and Mehl, 1933, Pa element; 03A-69-67, upper view. 14. *Amorphognathus* cf. *ordovicicus* Branson and Mehl, 1933, Pb element; 03A-69-72. (a) Outer oblique view; (b) upper view. 15. *Amorphognathus* cf. *ordovicicus* Branson and Mehl, 1933, Pb element; 03A-69-71. (a) Outer lateral view; (b) upper view. White drawings correspond to the basal sections. Scale bars = 200 μm.

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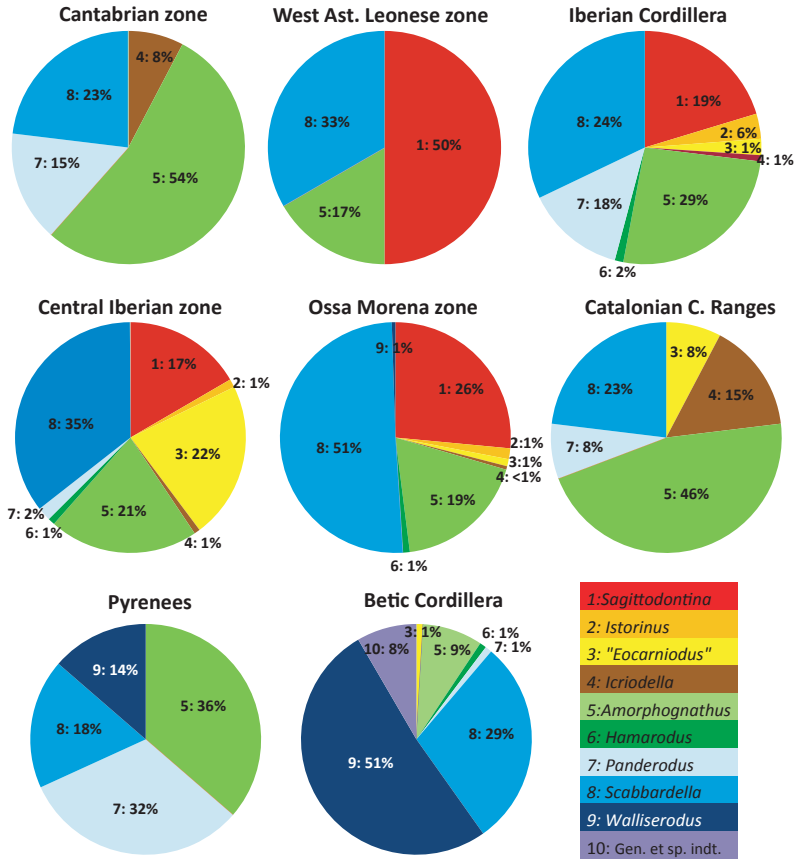

Fig. 6 Taxonomic compositions of Late Ordovician conodont taxa from different tectonostratigraphic zones of the Iberian Peninsula.

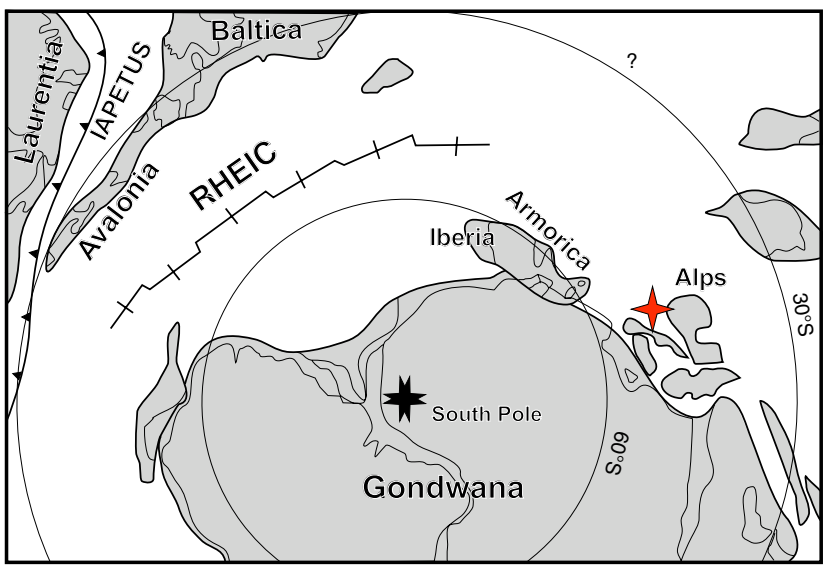

Fig. 7 Palaeogeographic reconstruction of North Gondwana during the Upper Ordovician (simplified from Cocks and Torsvik, 2006). The star indicates the proposed location of the Malaguide Complex of the Betic Cordillera 440 Ma ago.

Lindström (1976), Sweet and Bergström (1984) and Bergström (1990) located the Iberian Peninsula at very high palaeolatitude, as part of the Perigondwanic margin, during the latest Ordovician. This would explain quite well the absence of conodonts and the predominance of glaciomarine sediments during the Hirnantian in most Iberian areas. Nevertheless, the affinities between the Ardales Ordovician fauna and the conodont associations found in the Carnic Alps rather suggest a palaeogeographical location for the Malaguide Complex at lower palaeolatitude than the Iberian Massif, as proposed by Cocks and Torsvik (2002, 2006) for the Carnic Alps, or by Vennin *et al.* (1998) for the northeastern regions of Iberia. This palaeobiogeographical interpretation is geologically significant and reinforces a previous hypothesis supporting close palaeogeographical relationships between the Palaeozoic (and younger) successions of the Betic-Rifian Internal Domains and those of the Alps (see Chalouan and Michard, 1990; Michard *et al.*, 1991, among others).

Conclusions

1 The presence of Upper Ordovician rocks in the Malaguide Complex is demonstrated for the first time.
2 The Late Ordovician conodont fauna discovered at Ardales is correlated with the upper part of the *A. ordovicicus* Biozone, corresponding to the Hirnantian Stage. This is supported by: (i) its dissimilarity from typical Katian associations of the Mediterranean Province reported in other areas of the Iberian Massif; (ii) its similarity to associations recorded from the upper levels of the Uggwa limestones of the Carnic Alps and to the Hirnantian fauna of the Ashgill Series of Northern England and North Wales; and (iii) the presence of *Walliserodus* elements showing morphological affinities to Early Silurian species.
3 The conodont association found at Ardales indicates deposition at a lower palaeolatitude than that of other Iberian regions. It is similar to conodont faunas from the Carnic Alps, thus suggesting that close palaeogeographical relationships

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exist between the Malaguide Complex and the Alps.

4 Accordingly, during the Late Ordovician, the Malaguide Complex was located at an intermediate palaeogeographic position between the Carnic Alps and the Iberian terranes (Fig. 7).

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PRIMEROS DATOS SOBRE LA PRESENCIA DE CONODONTOS DEL EMSIENSE (DEVÓNICO INFERIOR) EN EL COMPLEJO MALÁGUIDE DE LA PROVINCIA DE GRANADA

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Resumen

El Paleozoico de la Cordillera Bética aflora casi exclusivamente en las Zonas Internas y suele estar formado por rocas metamórficas en la mayoría de los casos. Por ello, las dataciones en estos terrenos son muy escasas y, en su mayor parte, se han obtenido en los afloramientos occidentales del Complejo Maláguide. Con la presente nota se da a conocer el primer nivel fosilífero con conodontos del Paleozoico maláguide en su sector central (Zona de Cogollos Vega). Este nivel de calizas ha proporcionado la primera datación con conodontos del Emsiense (Devónico Inferior) de la provincia de Granada. Pese a que no se han encontrado niveles de calizas alabeadas en los alrededores, dado que este banco se encuentra en serie y no es un bloque redepositado, debe ocupar una posición estratigráfica equivalente a alguno de los niveles calcáreos de la parte intermedia de la Formación Santi Petri.

Palabras clave: conodontos, Emsiense, Complejo Maláguide

Abstract

The Paleozoic terrains of the Betic Cordillera crop out almost exclusively in the Internal Domain. They are mostly made up of metamorphic rocks. Therefore time-controls by means of fossils are scarce and they have been obtained mainly in the western outcrops of the Malaguide Complex. In this paper we report the first bed with conodonts of the central Malaguide Complex (Cogollos Vega Zone). This limestone level yields the first conodonts of Emsian age (Lower Devonian) in the province of Granada. Although we have not found *calizas alabeadas* in the surrounding area and because this level crops out in sequence and it is not a reworked block, it could be stratigraphically equivalent to any of the calcareous beds that are common in the middle parts of the Santi Petri Formation.

Keywords: conodonts, Emsian, Malaguide Complex

Introducción

El Complejo Maláguide constituye la unidad tectónica superior de la pila de mantos de corrimiento que conforma las Zonas Internas de la Cordillera Bética. Aflora desde las proximidades del Estrecho de Gibraltar hasta la provincia de Murcia, coincidiendo con el contacto entre las Zonas Internas y las Zonas Externas (FIG. 1A).

En la serie estratigráfica del Complejo Maláguide predominan los terrenos del Paleozoico (Grupo Piar: Martín-Algarra, 1987), que son recubiertos en discordancia por los del Triásico (Formación Saladilla: Roep, 1972). La estratigrafía de detalle del Paleozoico maláguide es aún poco conocida, en parte a causa de la complejidad de la estructura tectónica, pero también debido a la notable uniformidad de las facies terrígenas paleozoicas y a la discontinuidad lateral de las escasas y delgadas intercalaciones carbonatadas existentes. Éstas, además, muestran un variado espectro de facies, desde marinas someras a hemipelágicas y pelágicas, y han proporcionado fósiles de diversos horizontes del Silúrico a la base del Carbonífero Superior, a menudo en forma de clastos redepositados en formaciones conglomeráticas paleozoicas (Martín-Algarra et al., 2004).

El objetivo de la presente nota es dar a conocer el primer hallazgo de conodontos en el Complejo Maláguide de la provincia de Granada, lo que permite una primera correlación del Paleozoico de este sector con las sucesiones mejor conocidas de otros sectores.

Estratigrafía del Paleozoico maláguide en la Zona de Cogollos Vega

El complejo llamado originalmente "Bético de Málaga", definido en el sector occidental de la Cordillera Bética (Blumenthal, 1927) y posteriormente denominado Complejo Maláguide (Durand-Delga, 1968), fue reconocido por primera vez en la región al NE de Granada por Blumenthal (1928). Este complejo aflora a lo largo de una franja de terrenos predominantemente paleozoicos y triásicos que Blumenthal y Fallot (1935) denominaron Zona de Cogollos Vega (FIG. 1B) y que corresponde al afloramiento más extenso del Complejo Maláguide en la provincia de Granada.

Según García-Dueñas y Navarro-Vilá (1976) y Navarro-Vilà y García-Dueñas (1980) la litoestratigrafía de la sucesión paleozoica de la Zona de Cogollos Vega estaría definida por la presencia de las siguientes asociaciones de facies, que de más antigua a más moderna son: a) pizarras satinadas con niveles cuarcíticos y de conglomerados de cantos estirados; b) pelitas, grauvacas y conglomerados con intercalaciones de calizas alabeadas; y c) pelitas, grauvacas y conglomerados, que se pueden correlacionar, respectivamente, con las formaciones Morales, Santi Petri y Almogía del Grupo Piar (Martín-Algarra, 1987). Previamente, en las proximidades del Cortijo de Florencia, Foucault & Paquet (1971) identificaron horizontes silíceos negros (liditas) correlacionables con la Formación Falcoña (Herbig, 1983), que están intensamente replegados y aparecen recubiertos en discordancia por conglomerados y arenas rojas del Triásico, lo que les permitió demostrar que los terrenos del Paleozoico maláguide también fueron afectados por deformaciones hercínicas. Finalmente, Herbig (1984) reconoció además la existencia de diversos afloramientos del Conglomerado Marbella. Los clastos carbonatados de esta última formación han sido recientemente muestreados y estudiados de forma sistemática por Heekeren (2003), quien ha identificado en ellos diversos tipos de facies marinas someras y pelágicas, algunas de ellas con fósiles del Devónico y del Carbonífero.

Las observaciones que aquí se presentan han sido realizadas en la transversal del sector de Las Mimbres, al S y SW del Tajo del Despeñadero. En este sector los terrenos del Paleozoico están representados predominantemente por pizarras pardo-oliváceas con niveles delgados (centimétricos a decimétricos) de areniscas grauváquicas con estructuras sedimentarias que ponen de manifiesto su carácter turbidítico.

En el seno de estas pizarras oliváceas de grano fino que afloran a lo largo de una pequeña barranquera entre suelos a unos 500 m al WNW de la Fuente de los Potros, y al que se accede por la vereda que se dirige hacia la Acequia del Fardes desde el paraje conocido como las Mimbres, aparece un nivel carbonatado discontinuo, del cual una muestra recogida (06A-60) ha proporcionado la fauna de conodontos que a continuación se describe. Se trata de un paquete de calizas grises azuladas, de alrededor de un metro de espesor y constituido por varios estratos calcáreos separados por lechos pizarrosos. El banco aparece aislado entre pizarras oliváceas aunque en sus proximidades aparecen también restos derrubiados de liditas.

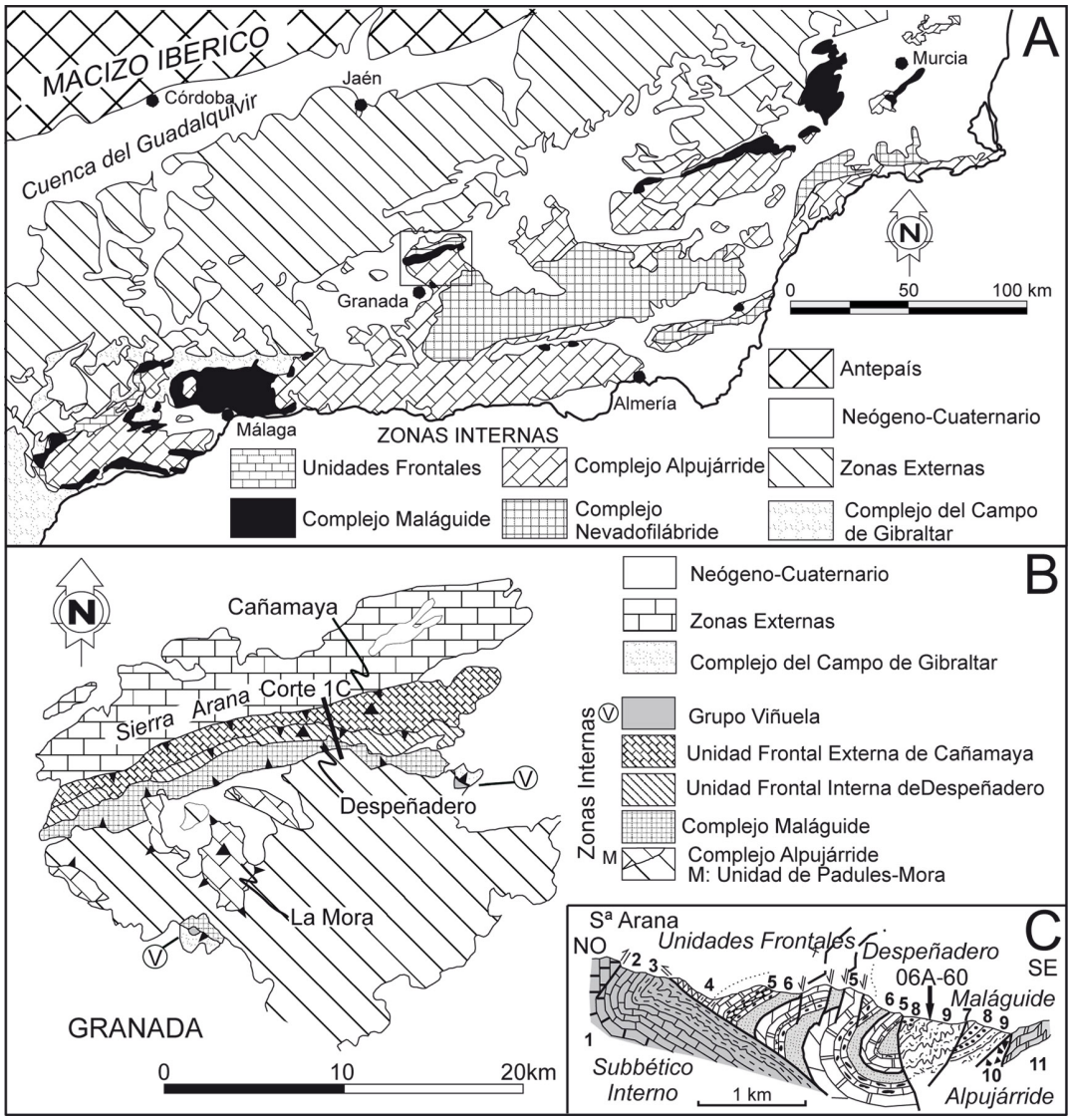

FIG. 1. A. Esquema geológico general de la Cordillera Bética. B. Esquema tectónico de la región al NE de Granada. La banda de Complejo Maláguide corresponde a la denominada Zona de Cogollos Vega. C. Corte geológico ilustrativo de la estructura del contacto entre las Zonas Internas y las Zonas Externas en el sector central de la Zona de Cogollos Vega.

Aunque su continuidad lateral no ha podido ser confirmada en las inmediaciones del afloramiento, y hasta la fecha no han podido ser observados niveles comparables en otros puntos de la Zona de Cogollos Vega, este nivel carbonatado corresponde con seguridad a un antiguo nivel estratigráfico intercalado entre la sucesión de pizarras, que ha sido severamente tectonizado, y puede excluirse que se trate de un bloque redepositado dentro de las pizarras.

Resultados

Se han disuelto 1,375 Kg de estas calizas (muestra 06A-60) mediante una disolución de ácido fórmico al 10%. La muestra ha proporcionado 4 elementos P_1 de *Criteriognathus steinhornensis* (Ziegler, 1956) y dos ejemplares coniformes de *Pseudooneotodus beckmanni* (Bischoff & Sannemann, 1958). *Criteriognathus steinhornensis* se caracteriza por presentar dos lóbulos laterales claramente asimétricos y con un contorno de redondeado a sub-cuadrado (FIG. 2). *Pseudooneotodus beckmanni* está representado por elementos coniformes con una cúspide recurvada y con un surco en la parte anterior de modo que en vista lateral se aprecia una pequeña joroba. La presencia de estas especies permite asignar este nivel al Emsiense, el último piso del Devónico Inferior.

Conclusiones

1) Se ha localizado el primer nivel fosilífero con conodontos del Paleozoico maláguide de la provincia de Granada, en la Zona de Cogollos Vega.

2) Este nivel ha proporcionado la primera datación con conodontos del Emsiense (Devónico Inferior) de la provincia de Granada.

3) Pese a que no se han encontrado niveles de calizas alabeadas en los alrededores, dado que este banco se encuentra en serie y no es un bloque redepositado, debe ocupar una posición estratigráfica equivalente a alguno de los niveles calcáreos de la parte intermedia de la Formación Santi Petri.

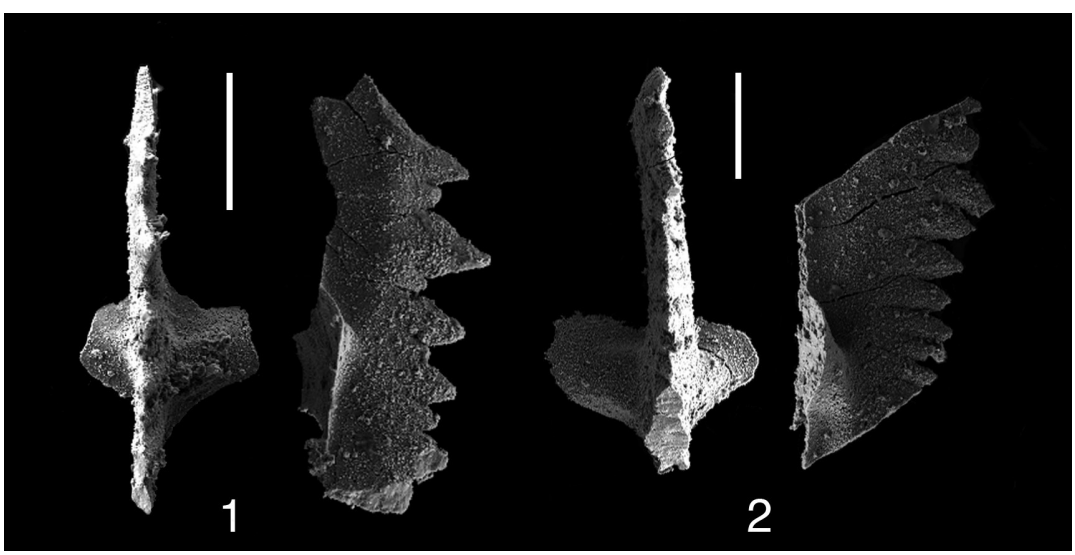

FIG. 2. Lámina de conodontos. Las escalas gráficas equivalen a 200 µm. 1. Elemento P_1 de *Criteriognathus steinhornensis* (Ziegler, 1956). Izquierda: vista superior; derecha: vista lateral. Ejemplar: 06A-60-4. 2. Elemento P_1 de *Criteriognathus steinhornensis* (Ziegler, 1956). Izquierda: vista superior; derecha: vista lateral. Ejemplar: 06A-60-1.

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Conodontos del Carbonífero en el Complejo Maláguide de la región de Ardales (Cordillera Bética occidental)

Carboniferous conodonts of the Malaguide Complex in the Ardales area (western Betic Cordillera)

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Palabras clave: Viseense, conodontos, Complejo Maláguide, Ardales, Formación Falcoña.
Keywords: Visean, conodonts, Malaguide Complex, Ardales, Falcoña Formation.

El Paleozoico de la Cordillera Bética aflora casí exclusivamente en las Zonas Internas. Está constituido fundamentalmente por terrenos metamórficos que no suelen contener fósiles y donde la estratigrafía original es difícil de establecer. Sin embargo, el Paleozoico del Complejo Maláguide, que es el complejo que ocupa la posición más alta en la pila de mantos que conforman las Zonas Internas, conserva gran parte de sus rasgos estratigráficos originales y es el único que permite hacer estudios estratigráficos y bioestratigráficos de detalle, lo que posibilita la reconstrucción de la historia geológica y paleogeográfica de este dominio. Pese al mal estado de conservación del contenido fósil de este complejo, se han podido datar distintos horizontes estratigráficos, sobre todo mediante conodontos, que van desde el Ordovícico Superior hasta el Carbonífero Superior (Rodríguez Cañero, 1993; Rodríguez-Cañero et al., 2010; Navas-Parejo et al., enviado, entre otros y referencias ahí).

La parte más baja del Carbonífero Inferior maláguide constituye la Formación Falcoña, definida por Herbig (1983). En esta formación se diferencian dos miembros, el inferior constituido por radiolaritas negras (liditas) datadas como Tournesiense mediante radiolarios, y el superior formado por calizas micríticas grises-azuladas, con conodontos del Viseense. Estas son facies típicas del Paleozoico de los dominios alpinos perimediterráneos y constituyen un horizonte de correlación muy importante a nivel regional, que ha

sido identificado en los Alpes, el Arco Cálabro-Peloritano, los Pirineos, Menorca y los Kabilides, entre otros lugares.

La Formación Falcoña, especialmente su miembro inferior radiolarítico, constituye un horizonte de correlación dentro del Complejo Maláguide. En este trabajo se presentan nuevos hallazgos de conodontos del Carbonífero Inferior en el área de Ardales (provincia de Málaga), que confirman la asignación de las liditas de esta área a la Formación Falcoña, y la presencia en ella de su miembro superior carbonatado.

El afloramiento estudiado se encuentra aproximadamente a 1 km al SO de Ardales, en el margen oriental del Arroyo de las Viñas (también llamado Arroyo de los Ángeles). En esta zona, buzando al norte, se observa un potente paquete de calizas turbidíticas en facies de *Calizas alabeadas*, perteneciente a la Formación Santi Petri, atribuida al Devónico. Sobre él descansa un tramo pelítico de unas decenas de metros de espesor, en el que localmente aparecen lentejones métricos de rocas subvolcánicas básicas, sobre el cual se observa una banda muy estrecha y lateralmente discontinua de liditas. Por encima de estos niveles, en un solo punto de la margen derecha del Arroyo de las Viñas y en contacto estratigráfico con unos decímetros de liditas, se ha observado una lente de calizas azuladas en plaquetas de menos de 1 metro de espesor. Las muestras tomadas en esta lente no han liberado restos de conodontos. Sin embargo, dos muestras tomadas en un bloque de unas calizas semejantes que aparece deslizado en el fondo del barranco y a pocos metros del afloramiento mencionado, han liberado fragmentos de elementos conodontales, muy deformados pero identificables, que indican una edad Carbonífero Inferior.

Entre esos fragmentos se han encontrado varios elementos Pa, que pese al mal estado de conservación, han podido identificarse como pertenecientes al género *Gnathodus*. Aunque están en parte fragmentados y muy mal conservados, el parapeto típico de este género se puede observar claramente. En el ejemplar A, ilustrado en la Figura 1, se conserva también parte de la cúpula externa y puede verse la fila de nódulos que se dispone paralela a la carena posterior. Esta característica identifica a *Gnathodus bilineatus bilineatus* (Roundy, 1926), de edad Viseense (Figura 1A). La edad carbonífera se confirma por la presencia de *Lochriea*

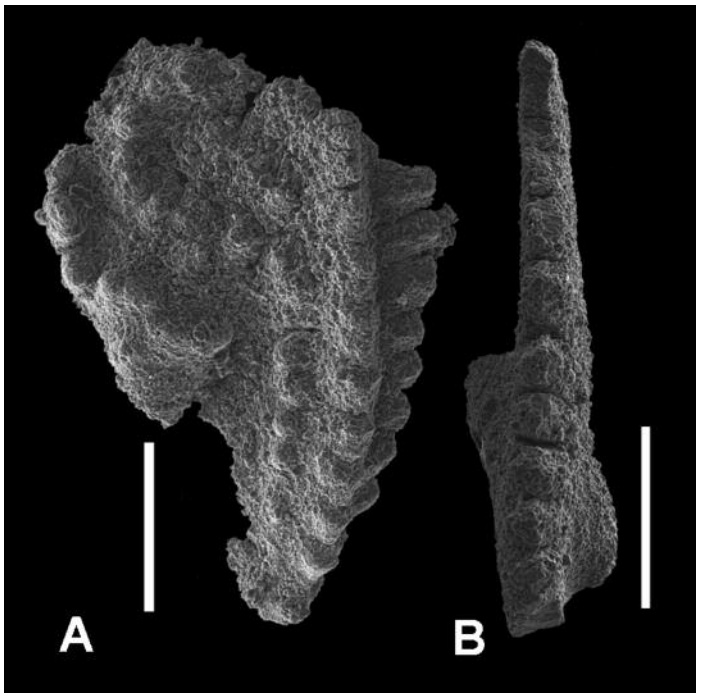

Figura 1. Conodontos viseenses en el área de Ardales. Todos son elementos Pa. Todas las escalas gráficas equivalen a 200 μm. A) *Gnathodus* cf. *bilineatus bilineatus*. Ejemplar 07P-20-4. B). *Lochriea commutata*. Ejemplar 07P-20-5.

commutata Branson & Mehl, 1941 en una de las muestras. Esta especie se caracteriza por elementos Pa carminiescafados con plataformas lisas y ligeramente asimétricas, cuyos dentículos anteriores y posteriores tienden a ensancharse lateralmente (Figura 1B).

Este hallazgo permite confirmar la edad carbonífera propuesta para las liditas fuertemente deformadas que aparecen a la base del Castillo de Ardales (Martín-Algarra et al., 2009), en el mismo cinturón cartográfico y en posición estratigráfica equivalente, con respecto a la Formación Santi Petri, a la del nivel aquí datado.

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Navas-Parejo, P., Rodríguez-Cañero, R. & Martín-Algarra, A. Enviado. Primer registro de un horizonte estratigráfico hemipelágico con conodontos del Carbonífero Superior en el Complejo Maláguide (Cordillera Bética oriental). *Geogaceta*, 52.

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Rodríguez-Cañero, R., Martín-Algarra, A., Sarmiento, G.N. & Navas-Parejo, P. 2010. First Late Ordovician conodont fauna in the Betic Cordillera (South Spain): a palaeobiogeographical contribution. *Terra Nova*, 22, 330-340.

Primer registro de un horizonte estratigráfico hemipelágico con conodontos del Carbonífero Superior en el Complejo Maláguide (Cordillera Bética oriental)

First record of a conodont-bearing, hemipelagic Upper Carboniferous stratigraphic horizon in the Malaguide Complex (eastern Betic Cordillera)

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ABSTRACT: A hemipelagic carbonate horizon interbedded with slates in the Malaguide Complex outcropping in the Vélez Rubio-Lorca Corridor provided a lowermost Bashkirian-middle Moscovian conodont association. This age is equivalent, or slightly younger, to the limestone boulders bearing shallow marine fossils found in the Marbella Conglomerate, which were the youngest beds found, up to now, in the Malaguide Paleozoic. This finding reveals that the Carboniferous palaeogeography in the Malaguide Domain was more complex than previously suspected, and that a lateral transition from shallow marine carbonate platforms to hemipelagic basinal environments persisted in this domain up to, at least, the late Carboniferous.

Key-words: *Malaguide Complex, Upper Carboniferous, hemipelagic limestones, conodonts.*

RESUMEN: Un horizonte carbonatado hemipelágico entre pizarras del Complejo Maláguide del Corredor de Vélez Rubio-Lorca ha proporcionado conodontos del Bashkiriense basal-Moscoviense medio. Esta edad es equivalente a (o ligeramente más moderna que) la de los clastos de calizas con fósiles marinos someros identificados en el Conglomerado de Marbella, que eran hasta la fecha los elementos más modernos reconocidos en el Paleozoico maláguide. Este hallazgo revela una paleogeografía carbonífera en el Dominio Maláguide algo más compleja de lo que hasta ahora se sospechaba, y pone de manifiesto la persistencia, en dicho dominio, de una zonación de ambientes de plataforma somera a cuenca hemipelágica hasta el Carbonífero tardío.

Palabras clave: *Complejo Maláguide, Carbonífero Superior, calizas hemipelágicas, conodontos.*

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Introducción

El Paleozoico de las Zonas Internas de la Cordillera Bética está sobre todo constituido por terrenos metamórficos cuya estratigrafía suele ser difícil de reconocer. No obstante, el Paleozoico del Complejo Maláguide conserva gran parte de sus rasgos originales, y suele estar formado por sedimentos de facies marinas profundas, turbidíticas y pelágicas. En ellas, el contenido fósil suele ser escaso y estar mal preservado pero, en algunas ocasiones permite realizar estudios estratigráficos detallados. Así, ha sido posible datar, mediante conodontos, niveles del Ordovícico Superior al Carbonífero Inferior. Estos estudios indican, además, que los rasgos fundamentales de la paleogeografía y evolución geohistórica del Paleozoico maláguide difieren de los observados en otras regiones de Iberia y los aproximan a los del Paleozoico de los Alpes (Herbig, 1985; Rodríguez-Cañero, 1993; Rodríguez-Cañero *et al.*, 2010, entre otros).

Los niveles más modernos del Paleozoico maláguide datados con fósiles corresponden al límite entre el Carbonífero Inferior y el Superior. Se han identificado en clastos de caliza incluidos en el Conglomerado de Marbella (Geel, 1973; Buchroithner *et al*., 1980; Herbig, 1985), que es la formación más moderna del Paleozoico maláguide. Se trata de calizas con foraminíferos bentónicos, corales, algas y otros fósiles de aguas marinas someras que llegan hasta el Bashkiriense basal.

En este trabajo se presenta el primer hallazgo de conodontos del Carbonífero Superior en el Complejo Maláguide. Estos aparecen en un horizonte carbonatado de facies hemipelágica, muy deformado e interestratificado con pizarras, que se ha localizado en el Corredor de Vélez Rubio-Lorca, el mayor afloramiento del Paleozoico maláguide en el sector oriental.

Contexto geológico

El Corredor de Vélez Rubio-Lorca es una depresión estrecha, en gran parte recubierta por terrenos postorogénicos del Plio-Cuaternario, que discurre por el contacto entre las Zonas Internas y las Zonas Externas de la Cordillera Bética entre las cuencas postorogénicas de Guadix-Baza (provincia de Granada) y de Lorca (provincia de Murcia). Limitado al N por relieves pertenecientes al Subbético (Sierras de Orce y María) y al S por terrenos del Complejo

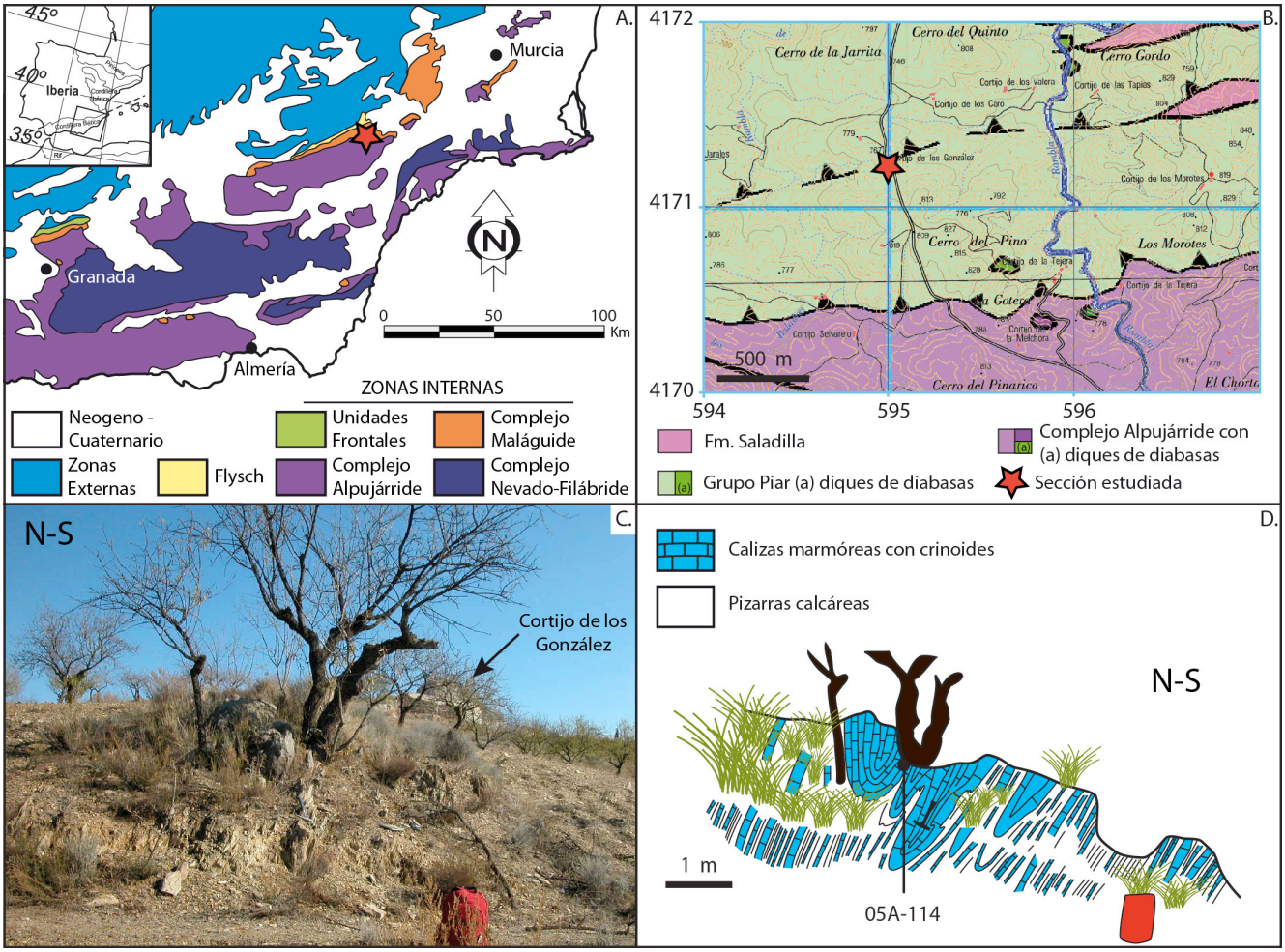

Fig. 1.- A. Esquema tectónico de la Cordillera Bética centro-oriental, con localización del afloramiento estudiado (estrella). B. Mapa geológico del Complejo Maláguide en el Cortijo de los González y su entorno (Mapa geológico de España, Plan Magna, hojas 952 y 974). C. Fotografía del afloramiento estudiado en 2007, antes de su destrucción. D. Esquema geológico del mismo afloramiento.

Fig. 1.- A. Tectonic sketch map of the central-eastern Betic Cordillera, with location of the studied outcrop (star). B. Geological map of the Maláguide Complex at the Cortijo de los González and neighbour areas (Geological map of Spain, Plan Magna, sheets 952 and 974). C. Field view of the studied oucrop in 2007 before its destruction. D. Geological sketch of the same outcrop.

Alpujárride (Sierra de las Estancias), a lo largo del corredor afloran ampliamente los terrenos del Complejo Maláguide, cuyas edades están comprendidas entre el Paleozoico inferior y el Mioceno Inferior. Estos terrenos se estructuran en varias unidades tectónicas alpinas, limitadas por contactos subverticales o buzantes al N, que separan escamas estrechas en número variable (generalmente de una a tres, a veces cuatro) según la transversal considerada.

La sección estudiada (Cortijo de los González; Fig. 1A) se encuentra al S de La Parroquia (provincia de Murcia) dentro de una unidad maláguide que Roep y Mac Gillavry (1962) denominaron Unidad de Casolidad. Roep (1972) puso de manifiesto que esta unidad se continúa hacia el E en la Unidad de la Peña Rubia de Lorca y que está constituida por terrenos del Paleozoico y del Triásico (Fig. 1B), afectados por una mayor deformación y recristalización metamórfica alpina que las que se observan en otras unidades maláguides más típicas de la misma región. Debido a ello, terrenos similares que afloran en la Cordillera Bética centro-occidental se han asignado al Complejo Alpujárride o bien a unas "unidades de transición" entre los complejos Maláguide y Alpujárride, (Sanz de Galdeano *et al*., 2001). Por esta razón, estos terrenos serán aquí incluidos en una *Unidad Maláguide Epimetamórfica*.

Las rocas paleozoicas de esta unidad se superponen tectónicamente a filitas grises azuladas y carbonatos del Triásico del Complejo Alpujárride de la Sierra de la Torrecilla. Por su parte, los terrenos triásicos de la Unidad Maláguide Epimetamórfica están tectónicamente recubiertos por una sucesión característica del Paleozoico maláguide, que Roep y Mac Gillavry (1962) asignaron a su Unidad de Cerro Colorado, la más baja tectónicamente, más extensa y potente del Complejo Maláguide en el Corredor de Vélez Rubio-Lorca. Por debajo de esta (habitualmente) *Unidad Maláguide Inferior*, la Unidad Maláguide Epimetamórfica ha sido tectónicamente laminada en muchos sectores, como puede observarse unos 2 km al oeste del afloramiento estudiado en el Cortijo de los González. De hecho, la Unidad Maláguide Epimetamórfica sólo aflora en el tercio oriental del Corredor de Vélez Rubio-Lorca.

El metamorfismo alpino que afecta a la Unidad Malaguide Epimetamórfica se constata claramente en las rocas detríticas triásicas, que aparecen transformadas a pelitas satinadas y filitas

rojas con niveles de cuarcitas rosadas y blancas. A pesar del metamorfismo que presentan, estas rocas triásicas son totalmente equivalentes a las de la Fm. Saladilla y, de hecho, Roep (1972) las incluyó en esta típica formación maláguide. Estas rocas afloran ampliamente en la Sierra de la Peña Rubia de Lorca, pero se estrechan hacia el O hasta desaparecer. En este caso, puede ser muy difícil separar los terrenos paleozoicos de la Unidad Maláguide Epimetamórfica de los que constituyen la base tectónica de las unidades maláguides suprayacentes.

El Paleozoico de la Unidad Maláguide Epimetamórfica está constituido por metapelitas y metapsammitas con niveles de microconglomerados. Son rocas muy parecidas a las que, en afloramientos maláguides más típicos, se asignan a la Fm. Morales, cuya edad se considera pre-devónica. Localmente, la sucesión incluye intercalaciones carbonatadas lateralmente discontinuas, de pocos metros de espesor, generalmente turbiditas calcáreo-pizarrosas de grano muy fino y facies semejantes a las calizas alabeadas de la Fm. Santi Petri. Esta última formación alcanza su máximo desarrollo en la Unidad Maláguide Inferior antes mencionada y es la más típica, potente y extensa del Devónico del Complejo Maláguide.

Corte del Cortijo de los González

El corte estudiado se encuentra en la carretera MU-701 de Henares a La Parroquia, a unos 10 km al norte de la salida 6 de la autovía A-91. Desafortunadamente, uno de los afloramientos, precisamente el datado en este trabajo e ilustrado en la Figura 1C-D, fue destruido en 2008 durante las obras de construcción de una tubería de gas.

Al N de la Venta de las Macetas, el corte de la carretera MU-701 permite observar cómo las filitas rosadas y plateadas del Triásico Alpujárride son recubiertas por las pizarras parduzcas que conforman, en sucesión ligeramente buzante al norte, la mayor parte de la serie paleozoica de la Unidad Maláguide Epimetamórfica. Esta serie paleozoica está cortada por diques de rocas subvolcánicas básicas semejantes a los que se observan en las rocas de la Fm. Morales en el área de Málaga. Asimismo, sus rasgos litológicos recuerdan mucho a los de los esquistos marrones de bajo grado del Complejo Alpujárride, que se observan bajo las filitas triásicas a lo largo de la misma carretera, así como a lo largo de la A-92N y alrededores.

Unos 300 metros al sur del Cortijo de los González, el contacto con el Alpujárride está bien expuesto y la sucesión comienza con alternancias de pizarras y metarenitas de grano fino, finamente estratificadas y de color marrón grisáceo que incluyen varias decenas de metros de calizas oscuras, finamente detríticas y bien estratificadas, semejantes a las facies de calizas alabeadas. El contacto entre estos materiales y los que afloran en el Cortijo de los González no se observa claramente, pero la misma sucesión de pelitas y metarenitas oscuras atravesadas por diques de diabasas puede verse, a lo largo de la carretera, al N del cortijo.

El corte aquí estudiado se ubica pocas decenas de metros al O del Cortijo de los González. Antes de su destrucción parcial constituía un lentejón carbonatado de unos pocos metros de espesor y un par de decenas de metros de continuidad lateral. Actualmente sólo pueden observarse escasos metros de este lentejón calcáreo, parcialmente cubiertos por suelos arados, y numerosos bloques calcáreos dispersos arrancados del mismo. Su extremo occidental, fotografiado en la Fig. 1C, cuya geología se ilustra en la Fig. 1D, y hoy destruido, permitía constatar que el lentejón carbonatado corresponde en realidad a una charnela sinclinal entre pizarras, muy apretada y cuyo eje es casi vertical. El núcleo del pliegue estaba formado por calizas micríticas grises-azuladas, de aspecto marmóreo, con estratificación media, que contienen artejos de crinoides a veces muy abundantes y que están rodeadas por pizarras calcáreas que alternan con niveles calcáreos deci- a centimétricos, asimismo fuertemente replegados. Una muestra de calizas recolectada en el núcleo del pliegue de este afloramiento (05A-114: Fig. 1D) ha proporcionado los primeros conodontos del Carbonífero Superior de la Cordillera Bética. Todavía es posible muestrear bloques aislados de estas calizas en el campo cultivado, pero ahora están fuera de su contexto estratigráfico y estructural, que antes, a pesar de su complejidad, era claramente visible.

Resultados bioestratigráficos

La disolución de 0.5 kg de muestra caliza con ácido acético al 10% tamponado con acetato cálcico, ha liberado 14 elementos Pa de conodontos y 59 elementos ramiformes y fragmentos, que indican una edad Carbonífero Superior (Bashkiriense basal a Moscoviense medio). Tres de los especímenes obtenidos más relevantes y mejor conservados se muestran en la Figura 2. Estos ejemplares se encuentran depositados en el Departamento de Estratigrafía y Paleontología de la Universidad de Granada.

La asociación faunística estudiada está formada por *Idiognathoides corrugatus* (Harris y Hollingsworth), *Declinognathodus noduliferous* (Ellison y Graves) y *Lochriea* aff. *commutata* (Branson y Mehl). El elemento Pa de *Idiognathoides corrugatus* se caracteriza por una plataforma ornamentada con costillas continuas sin surco medio y por una lámina que se une a la plataforma en el lado externo (Fig. 2A). Los elementos Pa asignados a *Lochriea* aff. *commutata* son carminiescafados, con plataformas lisas y ligeramente asimétricas, siendo estas significativamente más alargadas que en *Lochriea commutata* (Fig. 2B). Probablemente corresponden a formas transicionales que finalmente evolucionaron a *Declinognathodus orphanus* (van den Boogaard y Bless, 1985; von Bitter y Merrill, 1990).

El elemento Pa de *Declinognathodus noduliferous* presenta una lámina que se une a la plataforma centralmente, una pequeña carena que desciende hacia el parapeto externo y un surco medio somero, que se vuelve más ancho y profundo posteriormente. Los parapetos son paralelos y están ornamentados con nódulos (Fig. 2C).

Discusión y conclusiones

El hallazgo de un horizonte de calizas hemipelágicas con conodontos

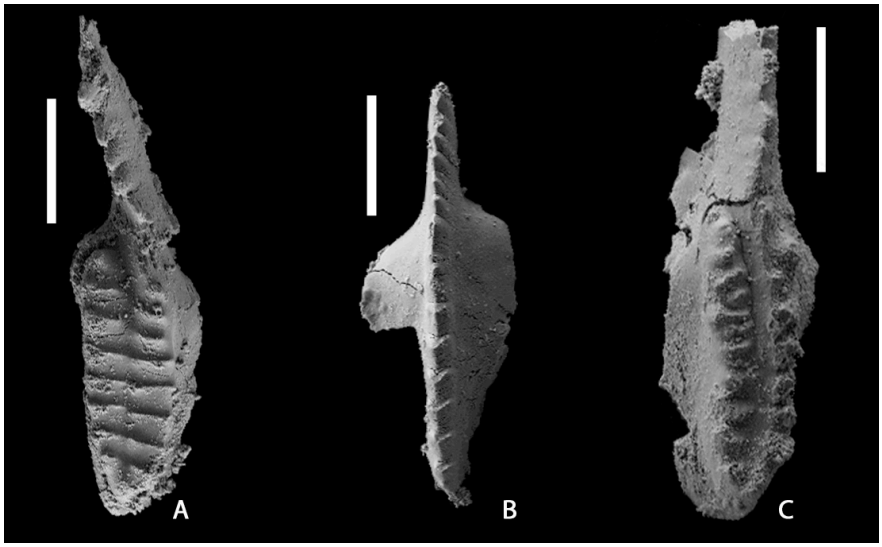

Fig. 2.- Conodontos del Carbonífero Superior del corte de los González. Las escalas gráficas equivalen a 200 µm. Todos son elementos Pa. A. *Idiognathoides corrugatus.* Ejemplar 05A-114-2. B. *Lochriea* aff. *commutata*. Ejemplar 05A-114-7. C. *Declinognathodus noduliferous*. Ejemplar 05A-114-4.

Fig. 2.- Upper Carboniferous conodonts of the González section. Scale bars are equivalent to 200 µm. All are Pa elements. A. Idiognathoides corrugatus*. Specimen 05A-114-2. B.* Lochriea aff. commutata*. Specimen 05A-114-7. C.* Declinognathous noduliferous*. Specimen 05A-114-4.*

en el Carbonífero Superior concuerda con lo observado en niveles más profundos de la serie paleozoica maláguide desde, al menos, el Ordovícico terminal, donde los fósiles que aparecen son casi exclusivamente propios de ambientes marinos abiertos típicos de la zona distal de un margen continental. Sin embargo, por su edad más moderna, contexto estratigráfico y litofacies, el horizonte carbonatado del Cortijo de los González difiere de otros identificados en la serie maláguide. En particular, el horizonte estudiado no está asociado a radiolaritas (liditas), a diferencia del otro nivel con conodontos del Carbonífero, datado del Viseense, que constituye el miembro superior de la Formación Falcoña (Navas-Parejo *et al*., 2008).

Por otra parte, a pesar de su notable deformación tectónica, el horizonte estudiado pasa gradualmente a la serie terrígena subyacente a través de una alternancia entre carbonatos y pelitas. Dicha serie terrígena muestra facies semejantes a las de las pizarras del miembro pelítico superior de la Fm Almogía, que incluye las facies Culm del Complejo Maláguide. Teniendo en cuenta lo anteriormente comentado, las calizas del horizonte aquí datado muestran facies hemipelágicas, con conodontos y fragmentos de crinoides pero sin otros fósiles de aguas someras, a diferencia de lo que se observa en los clastos de calizas del Bashkiriense incluidos en el Conglomerado de Marbella, en los que los únicos fósiles encontrados son propios de ambientes neríticos (corales, algas, foraminíferos bentónicos, entre otros organismos marinos del Carbonífero). Por tanto, el depósito de la sucesión estudiada en la zona distal de un margen continental paleozoico fue contemporáneo de una sedimentación en una plataforma carbonatada somera localizada en las regiones proximales de dicho margen.

Los nuevos datos obtenidos revelan que la paleogeografía maláguide durante el Carbonífero tardío fue algo más compleja de lo que hasta ahora se había sospechado, y demuestran la coexistencia de ambientes sedimentarios carbonatados neríticos y hemipelágicos (además de los ambientes turbidíticos siliciclásticos en los que se depositó la Fm. Almogía) hasta, por lo menos, el inicio del Carbonífero Superior. Un análisis más profundo de estos hechos, completado con nuevos hallazgos estratigráficos en niveles coetáneos y de otros más antiguos, permitirá profundizar en la zonación paleogeográfica del paleomargen representado por el Dominio Maláguide y ayudará a aclarar sus posibles relaciones con otros dominios paleozoicos ibéricos, norteafricanos y perimediterráneos.

Agradecimientos

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APPENDIX II

List of other contributions to journals and scientific meetings

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