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Terminal Cambrian and lowest Ordovician succession of Mexican West Gondwana: biotas and sequence stratigraphy of the Tiñu Formation

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Abstract – The Tiñu Formation of Oaxaca State is the only fossiliferous lower Palaeozoic unit between the Laurentian platform in northwest Mexico and Gondwanan successions in Andean South America. The Tiñu traditionally has been referred to the Lower Ordovician (Tremadoc) and regarded as having a provincially mixed fauna with Laurentian, Avalonian, and Gondwanan elements. Bio- and lithostratigraphic re-evaluation demonstrates that the Tiñu is a Gondwanan, passive margin succession. It includes a lower, thin (to 16 m), condensed, uppermost Cambrian Yudachica Member (new). The Yudachica nonconformably overlies middle Proterozoic basement as a result of very high late Late Cambrian eustatic levels. The Yudachica changes from storm-dominated, but slightly dysoxic, shelf facies (fossil hash limestone and shale) in the south to an upper slope facies with debris flows 50 km to the north. Three biostratigraphically distinct depositional sequences comprise the Yudachica. The Yudachica has Gondwanan-aspect trilobites with low-diversity conodonts characteristic of unrestricted marine/temperate facies. The upper Tiñu, or Río Salinas Member (new), is a Lower Ordovician (Tremadoc) depositional sequence that records strong early, but not earliest, Tremadoc eustatic rise marked by graptolite- and olenid-bearing dysoxic mudstones. Higher strata shoal upward into shell-hash limestones and proximal tempestite sandstones with upper lower Tremadocian unrestricted marine/temperate conodonts. New taxa include *Orminskia rexroadae* Landing gen. et sp. nov. from the *Cordylodus andresi* Zone; this euconodont is related to hyaline coniform genera best known from Ordovician tropical platform successions. *Cornuodus*? *clarkei* Landing sp. nov. resembles the coeval, upper lower Tremadoc tropical species *Scalpellodus longipinnatus* (Ji & Barnes).

Keywords: Cambrian, Ordovician, Gondwana, Mexico, conodonts, sequence stratigraphy.

1. Introduction

The Tiñu Formation in Oaxaca State, southern Mexico, provides the only record of fossiliferous lower Palaeozoic rocks south of the Laurentian successions in Sonora and Chihuahua states and Texas and north of the Gondwanan sequences in Andean Columbia and Venezuela. Pantoja-Alor (1970) discovered the small inliers of the Tiñu Formation in 1964 (Fig. 1). Their first report by Pantoja-Alor & Robison (1967) was followed by descriptions of trilobites and molluscs from the inliers (Robison & Pantoja-Alor, 1968; Yochelson, 1968; Flower, 1968). However, questions developed about the biostratigraphic and palaeogeographic significance of the Tiñu faunas, their depositional environments, and the early Palaeozoic location of southern Mexico.

A revision of the Tiñu trilobite fauna, based on new material, will be presented elsewhere, but preliminary assessments are included in this paper. Robison & Pantoja-Alor (1968) regarded the Tiñu trilobites as indicating correlation with lowest Ordovician (Tremadoc) faunas of the Avalon continent in England and Wales, as well as with the *Saukia* Zone of the classic Upper Cambrian succession of Laurentia. In particular, they suggested that the association of *Saukia globosa* Robison & Pantoja-Alor (now assigned to *Mictosaukia* Shergold, 1975) and *Richardsonella variagranula* Robison & Pantoja-Alor (variously assigned to *Pseudokainella* Harrington, 1938 or *Elkanaspis* Ludvigsen, 1982; see Ludvigsen, 1982; Jell, 1985; Peng, 1990*b*) in the lower Tiñu Formation corroborated Whittington's (1966, table 1; see also Wolfart, 1970, table 3) view that the Laurentian *Saukia* Zone was correlative with the Avalonian lower Tremadoc. Support for this latter correlation waned quickly (e.g. Ludvigsen, 1982), but some workers regarded at least the lower part of the Tiñu as Upper Cambrian. Thus, similarities in *Shumardia* species led Rushton (1982) to propose that the lower Tiñu correlated with the uppermost Cambrian *Acerocare*

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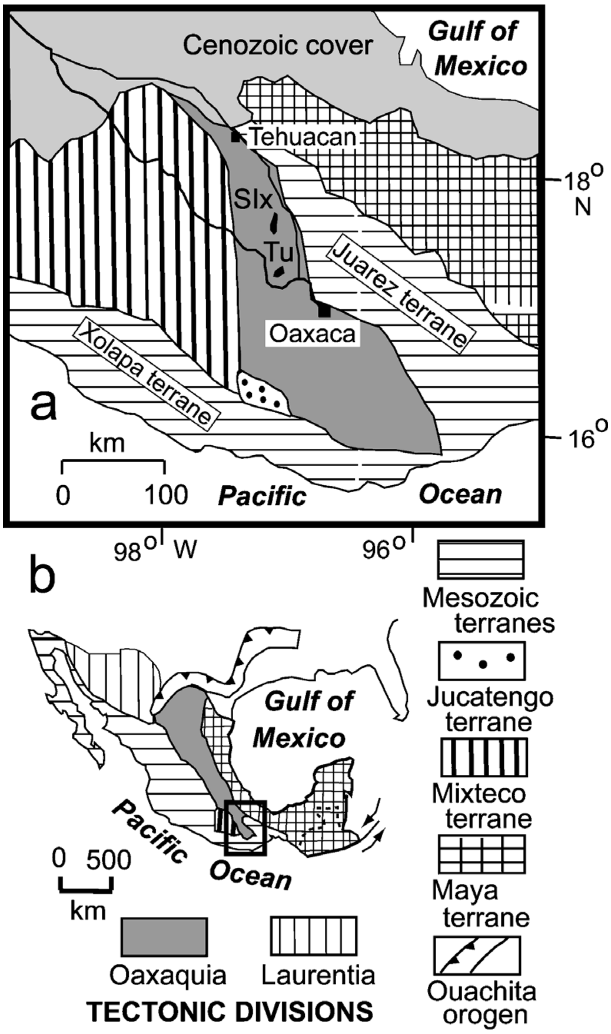

Figure 1. Generalized locality maps. (a) Terminal Cambrian and lowest Ordovician outliers of the Tiñu Formation (tiny black areas) at the Tiñu (Tu) and Barranco de Santiago Ixtaltepec (SIx) inliers on the Oaxaca Complex, Oaxaca State. Mesozoic terranes include the Xolapa and mylonitic Juarez terranes. (b) Major tectonic subdivisions of Mexico, Belize and northern Guatemala and relationship of Oaxachia to Ouachita orogen in northwest Mexico and adjacent United States. Area of (a) outlined by box in southern Mexico (modified from Centeno-Garcia & Keppie, 1999, fig. 1).

Zone of North Wales. In addition, Shergold (1988) emphasized *Neoparabolina frequens* (Barrande, 1868) (= *Parabolina argentina* (Kayser, 1876); see Nikolaisen & Henningsmoen, 1985) as a key to correlation with British, Scandinavian and Argentinian successions, and implied an assignment to the Upper Cambrian, rather than the Tremadoc (Shergold, 1988, fig. 2). Shergold (1988) also noted the significance of *Mictosaukia*, which indicates a correlation of the lower Tiñu Formation with pre-Tremadocian strata of South and North China (Peng, 1984; Qian, 1986).

However, these trilobite-based biostratigraphic re-evaluations ran counter to D. L. Clark's (*in* Robison & Pantoja-Alor, 1968, figs 2B–4) unillustrated reports from the lower Tiñu of the characteristic Tremadocian conodont *Cordylodus angulatus* Pander, 1856, and of *Oneotodus simplex* (Furnish, 1938), a species known from Laurentian Tremadoc-equivalent strata (Ethington & Brand, 1981). Recovery of nematophorous graptolites conclusively demonstrated that shales somewhat higher in the Tiñu at the Santiago Ixtaltepec locality (Fig. 1a, locality SIx) are lower Tremadoc (Sour & Buitrón, 1987; Sour Tovar, 1990), but left open the precise correlation of the lower Tiñu. Finally, even the depositional environment of the Tiñu Formation remained problematical, with biotic content (abundant echinoderm debris and associated green algal fragments) suggesting shallow, apparently photic marine conditions (Pantoja-Alor, 1970; Cabaleri & Armella, 1990). However, lenticular debris flows and interpreted deeper-water environments have been claimed to occur at all localities (Centeno-Garcia *et al.* 1998).

To resolve questions about the age, palaeobiogeographic significance and depositional environments of the Tiñu Formation, we undertook a restudy of the unit's conodont biostratigraphy and lithofacies. One likely consequence of this study was thought to be a refining of the early Palaeozoic location of southern Mexico, a region that has ping-ponged from proposed proximities with the northern Andes (Keppie, 1977), southern Laurentia (Scotese & McKerrow, 1990; Sedlock, Ortega-Gutiérrez & Speed, 1993), southern Andes (Dalziel, Dalla Salda & Gahagan, 1994), and Amazon craton (Keppie & Ortega-Gutiérrez, 1995; Keppie, 2004). Only much later in the Carboniferous did Oaxaquia, a continental fragment that comprises the middle Proterozoic basement and lower Palaeozoic cover sequence of central and southern Mexico (Fig. 1b), collide and unite with southern Laurentia to uplift the Ouachita orogen of south-central North America (e.g. Keppie *et al.* 2003; Keppie, 2004).

2. Geological setting and collecting localities

The Tiñu Formation nonconformably overlies middle Proterozoic (*c.* 1.0 Ga) high-grade metamorphics and intrusives of the Oaxacan Complex in two small inliers located northwest of Oaxaca City (Fig. 1a). Earlier overviews (Pantoja-Alor, 1970; Cabaleri & Armella, 1990; Centeno *et al.* 1998) documented marine depositional environments through the thin Tiñu Formation (72+ m: this report). Initial epsilon neodymium (εNd_i) and model ages (T_{DM}) from the Tiñu Formation are consistent with a sediment source from the Oaxacan Complex (Centeno-Garcia *et al.* 1998; Murphy *et al.* 2005; Gillis *et al.* 2005) and with early Palaeozoic passive margin deposition.

2.a. Locality Tu

At its type locality, the Río Salinas section of Pantoja-Alor (1970), the Tiñu Formation occurs in a

NNE-trending and −plunging syncline that is unconformably overlain on its east flank either by Cretaceous limestone or Tertiary red beds (see Centeno-Garcia & Keppie, 1999, fig. 3, for a detailed map of the type locality). The locality is 59 km NW of Oaxaca City, and is reached by exiting Federal Route MEX135 at Nochixtlán, driving 10 km S on Route 190, and then driving 2 km W on the dirt road to Río Salinas village. We measured a N-dipping, structurally simple, 72 m thick section through the Tiñu Formation in a heavily grazed goat pasture just east of the synclinal axis at the formation's type locality (Figs 1–3, locality Tu). Our locality Tu corresponds to Robison & Pantoja-Alor's (1968, fig. 2A, B) locality west of Tiñu village. This section extends from a contact with middle Proterozoic gneiss south of the dirt road and up the grassy slope to the highest exposures of the Tiñu Formation on the north shoulder of the road.

2.b. Locality SIx

Northern Tiñu Formation sections occur along the N–S-trending, W-dipping Barranco de Santiago Ixtaltepec inlier (Fig. 1a). This inlier nonconformably overlies the middle Proterozoic Oaxaca Complex. The Tiñu Formation in the Barranco de Santiago Ixtaltepec inlier is not unconformably overlain by Carboniferous marine rocks (compare Robison & Pantoja-Alor, 1968). Rather, the inlier is overlain to the west by a thrust slice of marine siliciclastics (Santiago Formation; Lower Carboniferous) and a higher slice of mixed marine siliciclastics and carbonates (Ixtaltepec Formation; Upper Carboniferous) (see Centeno-Garcia & Keppie, 1999, fig. 2, for a detailed map of the Barranco de Santiago Ixtaltepec inlier).

We did not re-investigate Robison & Pantoja-Alor's (1968, fig. 3) Arroyo Totoyac locality at the south end of the inlier, but measured and collected their northern section. This section is 66 km NW of Oaxaca City, and is reached by driving on a dirt road for 15.5 km NE of Nochixtlán village. The measured section lies 150 m past the last hairpin curve north of Santiago Ixtaltepec hamlet. It is reached by descending the steep path over the Carboniferous Ixtaltepec and Santiago formations 700 m north of Santiago Ixtaltepec (Figs 1–3; locality SIx). The lower 18 m of the Tiñu Formation at locality SIx is a monoclinal, W-dipping section, but higher shales and sheared fine-grained sandstones are increasingly faulted, folded, intruded with quartz veins and hydrothermally bleached toward the thrust contact with the coarse-grained, cliff-forming Santiago Formation.

3. Lithofacies and depositional environments

3.a. Members of the Tiñu Formation

As noted by Pantoja-Alor (1970; also Robison & Pantoja-Alor, 1968), the Tiñu Formation is divisible into a limestone-rich lower member and a shale- and sandstone-dominated upper member. We now distinguish these as the Yudachica Member (new) and Río Salinas Member (new), respectively (see Appendix 1). The base of the Río Salinas is a thin (to 15 cm), lenticular quartz arenite overlain by an equally thin, flat pebble conglomerate composed of limestone and silt-shale clasts derived from the Yudachica Member (Fig. 3).

3.b. Yudachica Member

The lower member of the Tiñu Formation consists of interbedded medium-grey limestones and light grey-brown, yellow-green weathering silt shales (Fig. 2). Up to 0.7 m of erosional relief on the Oaxacan Complex was observed at locality SIx, but abundant coarse sand derived from the Oaxacan Complex was observed only in the lower 0.85 m of locality Tu.

Diagenetic (nodular limestone and dolostone) carbonate horizons are rare in the Yudachica Member at locality Tu, but more common at SIx (Fig. 2). The dominant limestones at localities Tu and SIx range from minor calcisiltites, which are more abundant at SIx, to common fossil hash-intraclast wacke- to grainstones at both localities. The fossil allochems include disarticulated and fragmented trilobite debris, orthid brachiopods, and locally abundant gastropod conchs (particularly Tu-6.1, -6.2; SIx-6.5). Echinoderm ossicles are ubiquitous in the coarser limestones, and locally form grainstones (Tu-8.4–9.8). This abundance of fossil debris from a variety of benthic taxa, including rare calcified algae, in the coarser-grained limestones has been noted as evidence for lower Tiñu Formation deposition under relatively shallow, photic, unrestricted marine conditions (Cabaleri & Armella, 1990). The absence of carbonate ooids and aggregates, peloids, and any evidence of evaporates is consistent with unrestricted marine deposition, while the abundance of trilobite debris in almost all limestone beds comports with subtidal settings (e.g. Westrop & Adrain, 1998; Adrain *et al.* 2000). The combination of an absence of carbonate ooids or evaporates in limestones dominated by 'heterozoan' (particularly trilobite) fragments, and which lack obvious precipitated lime mud or marine cement (Cabaleri & Armella, 1990; thin-sections and slabs made in this study) are consistent with cool-water/temperate depositional regimes (e.g. James, 1997).

A common association noted in the field and in sawn slabs is that thin planar-laminated calcisiltite or trilobite wackestone beds have erosive tops when they are directly overlain by 5–10 cm thick, trilobite–echinoderm–calcisiltite intraclast pack- and grainstone beds with very thin carbonate sand caps. These calcisiltite/wackestone–pack-/grainstone relationships are more common at locality Tu, and were observed only in the lower Yudachica Member (lower 6.25 m at

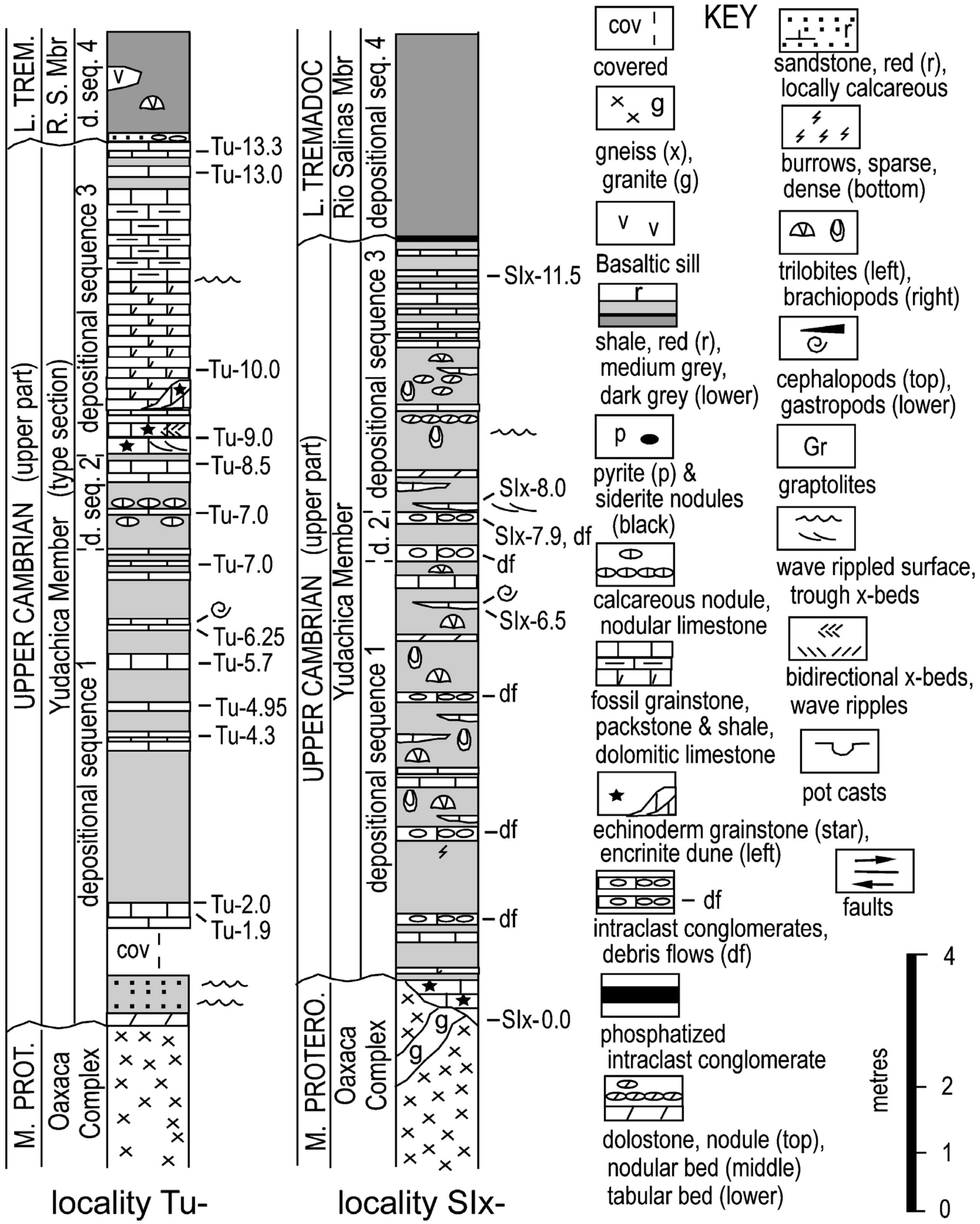

Figure 2. Sections in the lower Tiñu Formation (Yudachica Member); conodont samples (SIx and Tu) indicated to right of columns; number after locality abbreviation is metres above Oaxaca Complex. Figure shows proposed depositional sequences of Yudachica Member.

Tu and several horizons below 8.8 m at SIx). These relationships are interpreted to indicate condensation during limestone accumulation, and depositional processes that included erosion, winnowing and sorting under episodic high-energy conditions produced by storm waves.

Despite this evidence for episodic high-energy, wave-dominated deposition, microfacies and faunal evidence suggest the occurrence of relatively low oxygenation of bottom sediments and, possibly, bottom waters during Yudachica Member deposition. Formic acid-breakdown of limestones for acid-resistant microfossils yielded small amounts of polished chlorite (probably originally glauconite) sand grains and large quantities of phosphatic debris (up to 10 % of original sample mass). The phosphatic residue consists of a variety of components: rounded and polished, black phosphatic sand grains; abundant phosphatized conchs and steinkerns of gastropods, orthid brachiopods, hyoliths and ostracodes, some of which show the abrasion and polishing of already phosphatized ('pre-fossilized') grains before final burial; common phosphate-replaced agnostids and 'polymeroid' trilobite debris; and very large quantities of spongy, partially phosphatized fragments of intraclasts and lime mud. This phosphatization suggests replacement of calcium carbonate allochems and sediment with intense denitrification under an oxygen minimum layer or within dysoxic sediment before final burial, while the glauconite grains suggest proximity to a mixing zone between oxygenated and dysoxic/anoxic water masses (e.g. Sageman, Wignall & Kauffman, 1991). Evidence of low levels of bottom-water or sediment oxygenation is also suggested by the fact that intense burrowing activity is limited to burrow-churning in the wacke- and packstones, while the intervening grey-brown shales are either unburrowed or have only small *Planolites*-like horizontal burrows. Finally '*Olenus*'-like olenids and agnostids, trilobites characteristic of dark, organic-rich, low eH facies (e.g. Fortey & Cocks, 2003), appear through the Yudachica Member (Robison & Pantoja-Alor, 1968, figs 2–4; data from this study). Their presence through the Yudachica Member, but in association with abundant non-olenid trilobites (e.g. '*Asaphellus*'), echinoderms and brachiopods, suggests deposition of the Yudachica Member under water mass conditions that apparently ranged from weakly dysoxic to the lower end of 'normal' oxygenation (compare Schovsbo, 2001, fig. 3).

Primary sedimentary structures and depositional processes differ between localities Tu and SIx. Oscillatory wave activity is indicated at locality Tu by isolated sand ripples with bi-directional cross-beds (Tu-0.25–0.85), imbricated intraclast pebbles that form fans (Tu-7.1, -7.2), and a surface with small symmetrical ripples (Tu-11.3). Tidal action is also suggested at locality Tu by bi-directional troughs and climbing ripples in echinoderm grainstones (Tu-8.7–9.4 m) and a lensing echinoderm hash dune with west-dipping foresets (Tu-9.4 to Tu-9.8 locally).

Locality SIx differs by having a number of lenticular, intraclast conglomerates with particularly large (to 25 × 4 cm cross-sections), flat lime mudstone clasts in the lower 8.0 m of the Yudachica Member (Fig. 2; 1.7, 3.0, 5.1, 7.2, 7.9 m horizons). These conglomerates are debris flows, and the carbonate clasts generally 'float' in a muddy carbonate matrix with echinoderm fragments and sand- and granule-sized, lime mudstone fragments. In one bed (7.2 m), the flat limestone clasts are imbricated and dip to the south, and indicate transport to the north. Several normally graded, trilobite pack- to grainstone beds higher at locality SIx (8.0, 9.4 and 10.85 m) represent either turbidites or tempestites.

These observations agree with Centeno-Garcia *et al.*'s (1998) conclusion that debris flows occur in the lower Tiñu Formation. However, the debris flows occur to the north at locality SIx, while no evidence suggests that the locally intraclast-bearing pack- and grainstones at locality Tu reflect anything other than deposition on a storm wave-influenced shelf. The presence of debris flows, evidence for their northerly transport, and occurrence of turbidites/normally graded tempestite beds at SIx are all consistent with a northerly dipping ramp or slope during lower Tiñu Formation deposition.

3.c. Río Salinas Member

The sharp lithological contact of the Yudachica and Río Salinas members is marked by a thin, flat limestone and silt-shale clast conglomerate at localities Tu and SIx. The limestone clasts reach 30 × 20 × 7 cm at Tu, but are granule- and small pebble-sized at SIx, where the conglomerate is phosphatized and black in colour. The conglomerate and an underlying, lenticular green quartz arenite, which is present only at Tu, form the basal beds of the Río Salinas Member.

Limestones disappear at the Yudachica–Río Salinas contact, and the conglomerate is abruptly replaced by organic-rich, dark-grey, non-burrowed (laminated) shales with rare sideritic and pyrite nodules at Tu and SIx, respectively. This dysoxic facies comprises the lower 22.5 m of the Río Salinas Member at locality Tu. The dark shales have a low-diversity fauna that includes a few horizons with moldic olenid trilobites and nematophorous dendroid graptolites at localities Tu and SIx (Fig. 3).

A burrow-churned, nodular-bedded, trilobite wacke- to packstone with subangular phosphate pebbles overlain by a calcareous quartz arenite (Tu-35.7–36.0) marks the transition from dark-grey shale to medium-grey, structureless (burrow-churned), siliciclastic mudstone in the Río Salinas Member. (This limestone corresponds to Robison & Pantoja-Alor's (1968, fig. 2A) horizon 16000k in stratigraphic distance above the base of the Río Salinas Member, and is fault-duplicated in a flower structure on the west side of the syncline. Thus, we found that Robison & Pantoja-Alor's (1968, fig. 2A) trilobite-bearing horizons 16000i, 16000j are duplications of the 16000k bed.)

Above this lowest bedded limestone in the Río Salinas Member, the medium-grey siliciclastic mudstones have a number of limestones and calcareous, fine-grained sandstones, which are commonly

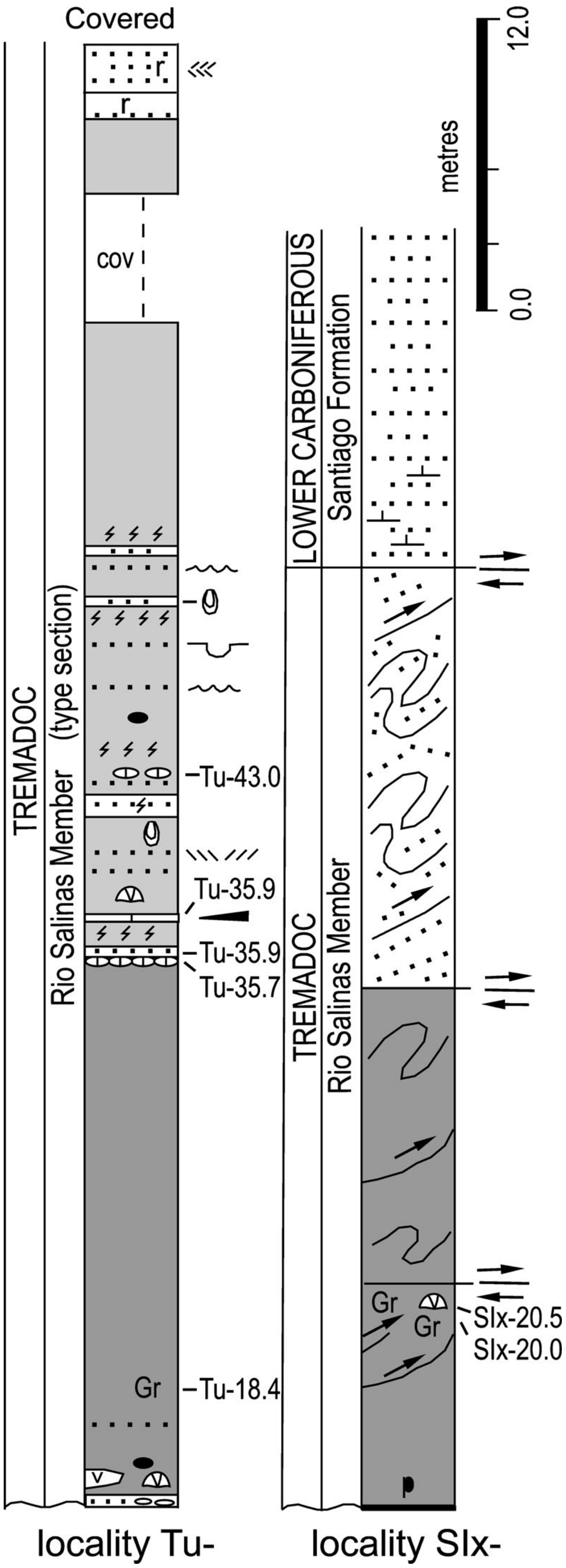

Figure 3. Sections in the upper Tiñu Formation (Río Salinas Member); samples (SIx and Tu) indicated to right of columns. Lithologies as in Figure 2.

burrow-mottled to -churned. Wave-produced sedimentary structures appear above this lowest bedded limestone, and include invaginated ellesmerceroid conchs (Tu-37.4; see Yochelson, 1968), bi-directional (wave) ripples (Tu-39.75), surfaces with symmetrical ripples (Tu-46.5, -51), and N–S-trending gutter casts (Tu-48.1). This interbedded shale and sandstone facies is suggestive of a shallow subtidal, proximal tempestite facies (e.g. Myrow, 1992). Improved oxygenation is suggested by the transition from dark-grey shales with sparse olenids and nematophorous dendroids to medium-grey shales with limestones that yield asaphids and remopleurids (horizon Tu-43). Still higher in the succession, the impression of a shoaling-up succession in the Río Salinas Member is further suggested by the presence of 3.2 m of red (that is, well-oxygenated) shale and coarse-grained sandstone with fragmented orthoid brachiopod shells and bi-directional, apparently tidally generated, dunes at the top of the exposure at locality Tu (Fig. 3).

4. Conodonts and biostratigraphy of Yudachica Member

4.a. Conodonts and colour alteration index

Conodont elements were not recovered from samples from the fossiliferous (e.g. trilobite and echinoderm hash) lowest beds of the Yudachica Member above the Proterozoic–Tiñu Formation nonconformity, and their correlation is problematical (Fig. 4). Abundant conodont elements appear about 2 m above the nonconformity, and persist through the top of the Yudachica Member (Table 1). The colour alteration indices (CAI) of conodont elements (see Epstein, Epstein & Harris, 1977) are low (CAI = 2) through most of the Yudachica Member at locality Tu, but abruptly increase to 3.5 in the upper 2 m, and this CAI is maintained in the Río Salinas Member at locality Tu. Colour alteration indices are much higher at locality SIx (CAI = 4), and this indicates deeper burial at Santiago Ixtaltepec.

4.b. *Cordylodus andresi* Zone

The lower 7.2 m of the Yudachica Member at Tu and the lower 6.5 m at SIx has a distinctive conodont fauna characterized by the oldest known cordylodan, *Cordylodus andresi* Viira & Sergeeva *in* Kaljo *et al.* 1986 (Fig. 5a–g). The definition of *C. andresi* followed herein is Szaniawski & Bengtson's (1998), with the species' elements having basal cavities that extend well above the zone of maximum cusp curvature, and sometimes almost reaching the tip of the element. As noted by Szaniawski & Bengtson (1998, p. 16), *C. primitivus* Bagnoli, Barnes & Stevens, 1987, *sensu* Müller & Hinz, 1991, from the Upper Cambrian of Sweden is best assigned to *C. andresi*. We further note that the type specimens and illustrated elements

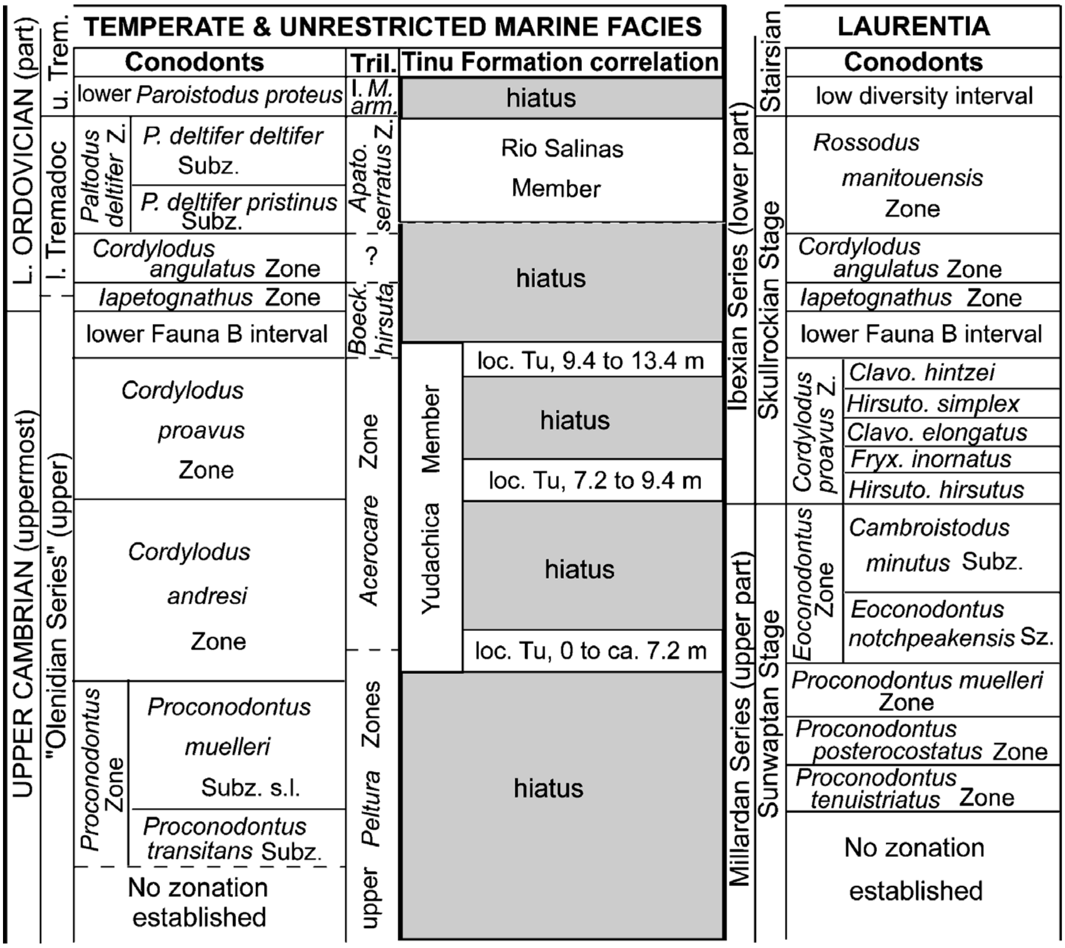

Figure 4. Conodont-based correlation of the Tiñu Formation into conodont and trilobite zonation of temperate successions of Baltica (left side of figure; see Szaniawski & Bengtson, 1998, and Löfgren, Repetski & Ethington, 1999) and into conodont succession of tropical Laurentia (e.g. Miller, 1980). Figure shows proposed depositional sequences of Yudachica Member.

of *C. primitivus* and *C. andresi* Barnes, 1988, have shallower basal cavities. As they fall within the range of morphology of *C. proavus* Müller, 1959, elements, the latter two ‘species’ should be assigned to *C. proavus* as junior synonyms.

Cordylodus andresi is reported only as low as the base of the *Cordylodus proavus* Zone on the Laurentian and North China carbonate platform (Miller *et al.* 2003; Dong, Repetski & Bergström, 2004; see Fig. 4). However, its occurrence is much lower in the Upper Cambrian of Baltica, where it is the eponymous species of a sub-*C. proavus* Zone (Szaniawski & Bengtson, 1998). This ‘low’ occurrence of *C. andresi* is also shown in the Yudachica Member (Tu-2.05, –4.95). As in southern Sweden (Müller & Hinz, 1991, table 2; Szaniawski & Bengtson, 1998), *C. andersi* occurs with *Proconodontus muelleri* Miller, 1969, and very weakly denticulated elements of *P. serratus* Miller, 1969 (Fig. 6n, o, t), species which have ranges that do not persist into the *C. proavus* Zone of Laurentia and North China. This lower part of the Yudachica Member is correlated with the Laurentian *P. muelleri* Zone (compare Szaniawski & Bengtson, 1998, p. 14).

Slightly higher in the Yudachica Member, the appearance of *Eoconodontus notchpeakensis* (Miller, 1969) (Fig. 6q, r) in Tu-5.7 and Tu-6.5 with all earlier-appearing *Cordylodus andresi* Zone taxa indicates a likely correlation with the *Eoconodontus* Zone of the Laurentian carbonate platform (e.g. Miller *et al.* 2003) (Fig. 4). *Teridontus nakamurai* (Nogami, 1967) (Fig. 6i–l) is common in the lower Yudachica Member (Table 1), but provides no further resolution in biostratigraphic correlation. Although *T. nakamurai* appears in sub-*Cordylodus proavus* Zone strata in unrestricted marine facies marginal to Laurentia (Landing, Ludvigsen & von Bitter, 1980; Landing, 1983), its lowest occurrence is in the lower *C. proavus* Zone on the Laurentian platform (Miller, 1980; Miller *et al.* 2003). *Orminskia rexroadae* Landing gen. et sp. nov. and species (Fig. 6a–h) is newly recognized from the Yudachica Member, and does not yet assist in interregional correlation. Its seeming disappearance within the *C. andresi* Zone after the appearance of *E. notchpeakensis* (Table 1) may be of importance in subsequent studies of the *C. andresi* Zone. The superficial resemblance of *O. rexroadae* gen. et sp.

Table 1. Upper Cambrian conodonts from productive samples from the Yudachica Member of the Tiñu Formation

	SIx -6.5	SIx -7.9	SIx -8.0	SIx -11.5	Tu -1.9	Tu -2.05	Tu -4.3	Tu -4.95	Tu -5.7	Tu -6.25	Tu -7.9	Tu -8.5	Tu -9.0	Tu -13.0	Tu -13.3
PROTOCONODONTS															
‘*Phakelodus elongatus*’									3	1		2	21	19	25
Phakelodus tenuis					1					1		2	6		3
EUCONODONTS															
Cordylodus andresi															
rounded (cordylodiform) el.	19							14	4	12					
compressed (cytroniodiform) el.	13														
Cordylodus caseyi															
rounded (symmetrical) el.			56	37									7	18	1
rounded (asymmetrical, ‘*caseyi*’) el.			22	12										6	
rounded (asymmetrical, ‘*drucei*’) el.			2	3										6	
compressed (cyrtoniodiform) el.			38	20									1	14	1
Cordylodus proavus															
rounded (cordylodiform) el.		12	35	59							15	33	33	35	1
compressed (cyrtoniodiform) el.		9	16	8							3	5	11	8	
Cordylodus prion sensu Nicoll (1991)															
rounded el.				6										3	
Cordylodus indet. (opaque black el.)															
rounded (cordylodiform) el.		4	56	20											
compressed (cyrtoniodiform) el.			12	9											
Eoconodontus notchpeakensis															
rounded (symmetrical) el.			5	10						8	4	5	25	1	1
compressed (cyrtoniodiform) el.		3										1	9		
Orminskia rexroadae gen. et sp. nov.						2	3	14	13	4					
Proconodontus muelleri						3		25		15					
Proconodontus serratus								2							
Semiacontiodus nogamii															
symmetrical (acontiodiform) el.															1
asymmetrical (unicostate) el.			2	1											1
Teridontus nakamurai			21	43	1				3			1	3	26	

Acid-disaggregated samples had a 6.0 kg mass. el. – element. Intense thermal blackening rendered some cordylodan elements from locality SIx- opaque; as the shape of their basal cavities cannot be determined, species-level identifications are not possible.

nov. and a possible assignment of its elements to *T. nakamurai* may explain reports of exceptionally low, sporadic occurrences of *Teridontus* Miller, 1980, on the Laurentian platform (Miller, 1980; Miller *et al.* 2003, fig. 5).

4.c. *Cordylodus proavus* Zone

Several samples in the middle Yudachica Member (Tu-7.9, -8.5; SIx-6.9) are assigned to an undivided *Cordylodus proavus* Zone. With the exception of *Eoconodontus nakamurai* and *Teridontus nakamurai*, all *Cordylodus andresi* Zone taxa disappear in the Yudachica Member with the appearance of *C. proavus* Müller, 1959 (Fig. 5h–l). None of the taxa used for subdivision of the *C. proavus* Zone into five subzones in Laurentian successions (Miller, 1969, 1980; Fig. 4) were recovered from this low-diversity (three euconodont species) interval. Thus, the *C. proavus* Zone of the middle Yudachica Member can only be considered as younger than the *C. andresi* Zone and older than the overlying terminal *C. proavus* Zone (Fig. 4).

4.d. Terminal *Cordylodus proavus* Zone–lower Fauna B interval

Horizons at least as low as Tu-9.0 and SIx-8.0 yield abundant elements of *Cordylodus caseyi* Druce & Jones, 1971, emend. Landing, 1993 (Fig. 5m–t), with rare specimens of *Semiacontiodus nogamii* Miller, 1969 (Fig. 6m, p). *Cordylodus caseyi* emend. is the senior synonym for numerous reports of *Cordylodus intermedius* Furnish, 1938, and several other cordylodan ‘species’ from the terminal Upper Cambrian, upper *C. proavus* Zone on a number of Cambrian palaeocontinents (see synonymy and discussion in Landing, 1993, pp. 13, 15, 16). An association of *C. caseyi* emend. and *S. nogamii* is present in the two upper subzones of the *C. proavus* Zone as originally defined by Miller (1969, 1980) (see discussion in Section 6.a) in Laurentia and China (e.g. Ross *et al.* 1997; Miller *et al.* 2003; Dong, Repetski & Ethington, 2004).

The low-diversity euconodont fauna from the upper Yudachica Member (five species) also includes a few elements of *Cordylodus prion* Lindström, 1955, *sensu* Nicoll, 1991 (Fig. 5u) recovered near the top of the Yudachica Member (Table 1; Tu-13.0 and SIx-11.5). Limited data exist on the stratigraphic range of *C. prion* in Laurentia and Australia (Nicoll, 1991; Miller *et al.* 2003). Its lowest occurrence in these tropical successions is in post-*Cordylodus proavus* Zone and terminal Upper Cambrian strata variously referred to the lower Fauna B interval (Ethington & Clark, 1971; Fig. 4) or *Cordylodus lindstromi* Zone (see Miller *et al.* 2003). Significantly, although *Cordylodus angulatus* has been reported from the

Figure 5. Upper Cambrian *Cordylodus* species from the Yudachica Member of the Tiñu Formation; hypotype specimens. (a–g) *Cordylodus andresi* Viira & Sergeeva *in* Kaljo *et al.* 1986, from Tu-4.95 unless otherwise indicated. (a) Rounded element, NYSM 17369, ×202; (b) rounded element, NYSM 17370, ×82; (c) rounded element, NYSM 17271, ×119; (d) rounded element, NYSM 17372, ×127, Tu-5.7 (see Fig. 10g); (e) rounded element, NYSM 17373, ×105; (f) compressed element, NYSM 17374, ×93 (see Fig. 10h); (g) compressed element, NYSM 17375, ×82. (h–l) *Cordylodus proavus* Müller, 1959. (h) Rounded element, NYSM 17376, ×98, Tu-9.8 (see Fig. 10i); (i) rounded element, NYSM 17377, ×87, Tu-13.0 (see Fig. 10j); (j) rounded element, NYSM 17378, ×69, Tu-8.5; (k) compressed element, NYSM 17379, ×69, Tu-9.8 (see Fig. 10k); (l) compressed element, NYSM 17380, ×84, Tu-9.8. (m–t) *Cordylodus caseyi* Druce & Jones, 1971, emend. Landing, 1993, from Tu-13.0 unless otherwise indicated. (m) Rounded element, NYSM 17381, ×79, Tu-9.8 (see Fig. 5m); (n) compressed element, NYSM 17382, ×62 (see Fig. 10p); (o) compressed element, NYSM 17383, ×65; (p) rounded (‘*Cordylodus drucei*’ *sensu formo*) element, NYSM 17384, ×57; (q) rounded element, NYSM 17385, ×78; (r) rounded element, NYSM 17386, ×69; (s) asymmetrical rounded element, NYSM 17387, ×78 (see Fig. 10n); (t) rounded element transitional into ‘*Cordylodus drucei*’ *sensu formo*, NYSM 17388, ×79. (u) *Cordylodus prion* Lindström, 1955, *sensu* Nicoll (1991), rounded element, NYSM 17389, ×116, Tu-13.0 (see Fig. 10q).

Yudachica Member (Robison & Pantoja-Alor, 1968), this diagnostic lower Tremadoc species (e.g. Löfgren, Repetski & Ethington, 1999) was not recovered in this study. In addition, Robison & Pantoja-Alor’s (1968) report of *Oneotodus simplex*, a species described from Laurentian Tremadoc-equivalent strata (Ethington & Brand, 1981), was also not confirmed in this study. The report of *O. simplex* from the Yudachica Member may have been based on elements of *Teridontus nakamurai.* In short, all available data emphasize that no faunal evidence indicates that the uppermost Yudachica Member is Tremadoc and lowest Ordovician.

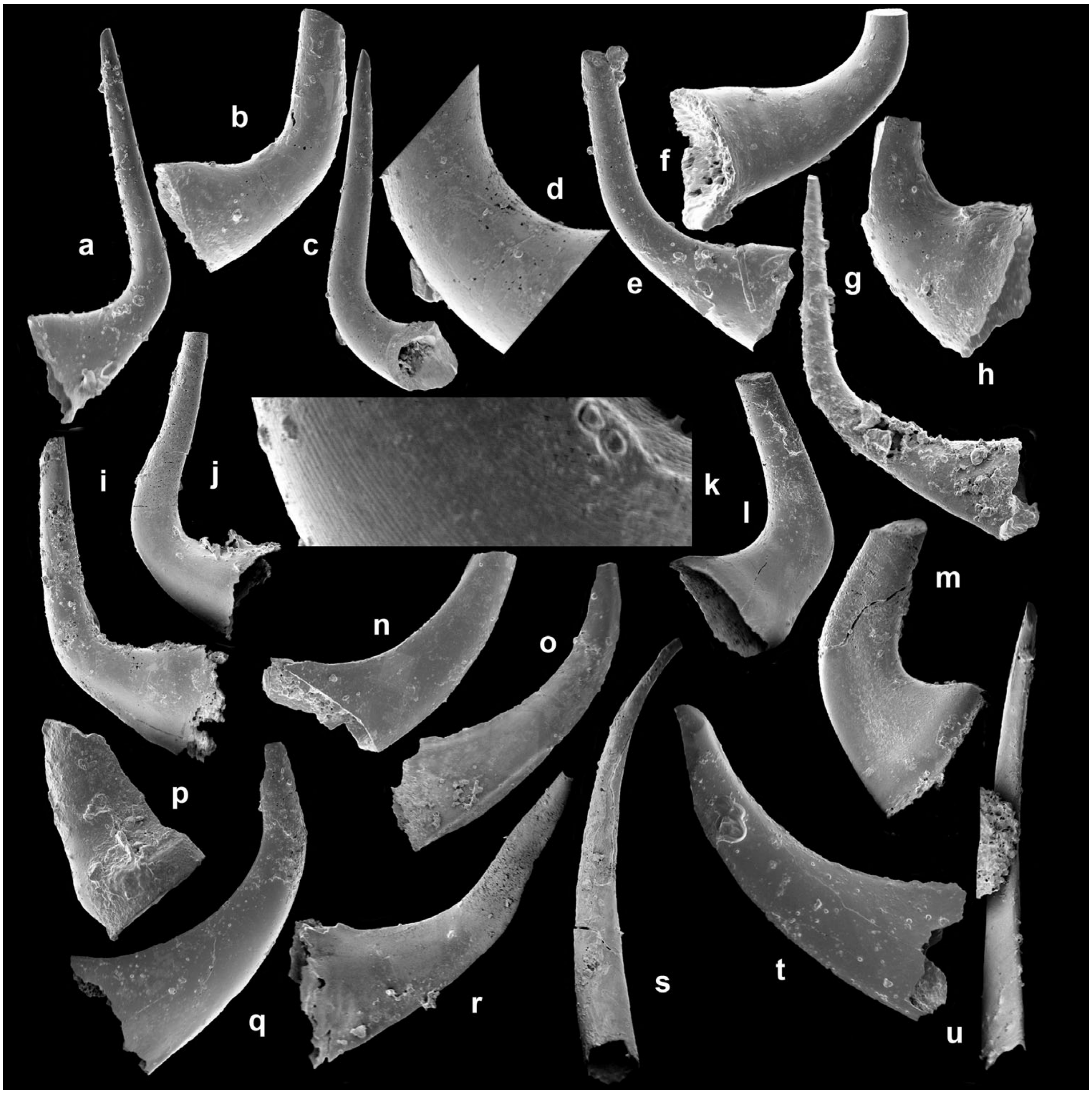

Figure 6. Upper Cambrian conodonts with coniform apparatuses from the Yudachica Member of the Tiñu Formation; hypotype specimens unless otherwise indicated. (a–h) *Orminskia rexroadae* Landing gen. et sp. nov. (a) Holoytpe NYSM 17390, ×108, locality Tu-2.05 (see Fig. 10a); (b) paratype NYSM 17391, ×178, Tu-2.05; (c, d) paratype NYSM 17392, lateral view ×126, and magnified view ×347 of smooth surface, Tu-4.3; (e) paratype NYSM 17393, ×160, Tu-4.95 (see Fig. 10b); (f) paratype NYSM 17394, ×198, Tu-4.95; (g) paratype NYSM 17395, ×212, Tu-4.95; (h) paratype NYSM 17396, ×174, Tu-5.7. (i–l) *Teridontus nakamurai* (Nogami, 1967). (i) NYSM 17397, ×85, Tu-5.7; (j, k) NYSM 17398, lateral view ×81 and detail of longitudinal microstriae ×775, Tu-9.8; (l) NYSM 17399, ×150, Tu-13.0. (m, p) *Semiacontiodus nogamii* Miller, 1969, from Tu-13.0. (m) Asymmetrical (anterolaterally weakly costate) element, NYSM 17400, ×154; (p) symmetrical (posteriorly costate) element, NYSM 17401, ×156. (n, o) *Proconodontus muelleri* Miller, 1969, from Tu-4.95. (n) Element with long oral edge (tip broken off during scanning microscopy), NYSM 15239, ×157; (o) NYSM 17402, ×127 (see Fig. 10d). (q, r) *Eoconodontus notchpeakensis* (Miller, 1969) from Tu-5.7. (q) Symmetrical (rounded) element, NYSM 17403, ×118 (see Fig. 10e); (r) asymmetrical (laterally deflected) element, NYSM 17404, ×144 (see Fig. 10f). (s) *Phakelodus tenuis* (Müller, 1959) from Tu-13.0; element with rounded cross-section, NYSM 17405, ×62. (t) *Proconodontus serratus* Miller, 1969; element with low serrations on oral margin, NYSM 17407, ×104, Tu-6.25 (see Fig. 10c). (u) '*Phakelodus elongatus*' (Zhang *in* An *et al.* 1983), element with sharp posterior margin, NYSM 17406, ×86.

5. Biostratigraphy of Río Salinas Member

5.a. Lowest Río Salinas Member

We collected nematophorous dendroid graptolites as low as 5.0 m above the base of the Río Salinas Member at locality Tu, and duplicated Sour & Buitrón's (1987) collections somewhat higher in the member at locality SIx (Fig. 3; Tu-18.0, SIx-22.0, -22.5). Only one form, *Rhabdinopora flabelliforme desmograptoides* (Hahn, 1912), is present in our collections (Fig. 7). Sour & Buitrón (1987) and Centeno-García *et al.* (1998, fig. 11.2) reported a larger dendroid fauna (six taxa) in

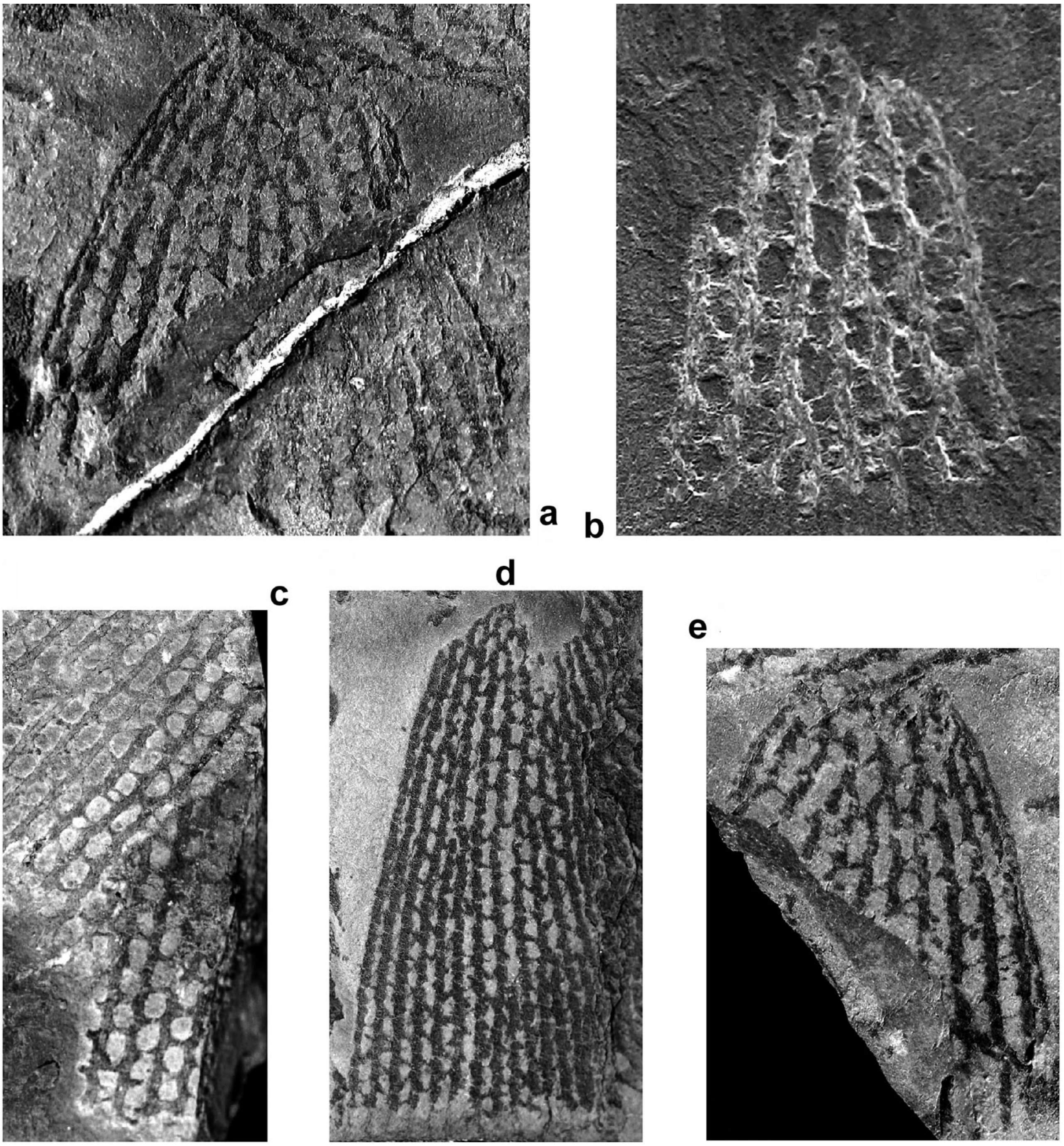

Figure 7. *Rhabdinopora flabelliforme desmograptoides* (Hahn, 1912) from the lower Río Salinas Member of the Tiñu Formation; hypotype specimens. (a, e) Part and counterpart, ×4.0 and ×3.8, of specimen showing branching in and shape of apical region of rhabdosome, NYSM 17408A and 17408B, SIx-22.0; (b) rhabdosome fragment in inverted view to emphasize stipes and dissepiments, NYSM 17409, ×7.1, SIx-22.0; (c) rhabdosome fragments with dissepiments that are thin (upper left) and ontogenetically thickened with thick bases (right), ×2.8, Tu-18.4; NYSM 17410; (d) rhabdosome fragment with gently convex left margin, elongate interstipe spaces, and stipe branching limited to apical region at top of figure, ×2.5, NYSM 17411, SIx-22.0.

the tectonically contorted lower Río Salinas Member at SIx. However, all of their forms are synonymized with *R. flabelliforme desmograptoides* (see Systematic Palaeontology).

This subspecies is known from cool-water successions on the Avalon and Baltic palaeocontinents (Hahn, 1912; Bulman, 1954), where it is an element of the middle lower Tremadoc graptolite ‘Assemblage 2’ (Cooper, 1979). Conodont-based correlations indicate that the Assemblage 2 ‘Matane graptolites’ of southern Quebec (see Bulman, 1950) are all referable to the *Rossodus manitouensis* Zone in Laurentia (Landing, Barnes & Stevens, 1986; Fig. 4). Consequently, the lower Río Salinas Member is also best correlated with

Figure 8. Lower Ordovician (upper lower Tremadocian) conodonts from the middle Río Salinas Member of the Tiñu Formation; hypotype specimens. (a–q) *Paltodus deltifer deltifer* (Lindström, 1955) from Tu-43.0. (a–e) oistodiform elements. (a) NYSM 17412, ×83; (b) NYSM 17413, ×78; (c) NYSM 17414, ×111; (d) NYSM 17415, ×82; (e) NYSM 17416, ×108. (f–i) Scandodiform elements. (f) NYSM 17417, ×63; (g) NYSM 17418, ×118; (h) NYSM 17419, ×75; (i) NYSM 17420, ×98. (j, k) Suberectiform elements, NYSM 17421 and NYSM 17422, ×108 and ×140; (l) symmetrical drepanodiform, NYSM 17423, ×109; (m) subsymmetrical drepanodiform with inner lateral sulcus on oral and aboral margins, NYSM 17424, ×159; (n) subsymmetrical drepanodiform with laterally deflected aboral keel elongated into anticusp, NYSM 15240, ×95; (o, p) subsymmetrical drepanodiform with weakly laterally deflected anterior keel, lateral view ×88 and detail of longitudinal microstriae ×330, NYSM 17425; (q) anterolaterally costate ('acodiform') element, NYSM 17426, ×122. (r–t) *Paroistodus numarcuatus* (Lindström, 1955) from Tu-35.9. (r) Oistodiform (M) element, NYSM 17427, ×118; (s, t) drepanodiform element, lateral view ×125 and enlarged view ×465 of longitudinal microstriae, NYSM 17428. (u–cc) *Variabiloconus variabilis* (Lindström, 1955) from Tu-35.9 unless otherwise indicated. (u–w) Symmetrical drepanodiforms; (u) NYSM 17429, ×86; (v, w) lateral view ×112 and enlarged view ×815 of longitudinal microstriae, NYSM 17430. (x) Weakly asymmetrical drepanodiform element with gentle anterolateral flare of aboral margin, NYSM 17431, ×89; (y) asymmetrical drepanodiform element with laterally displaced sulcus on oral margin, NYSM 17432, ×95; (z–bb) acontiodiform elements from Tu-43.0 with weakly developed anterolateral costae, NYSM 17433–17435, ×103, ×74 and ×93, respectively; (cc) asymmetrical element with anterolateral costa, NYSM 17436, ×80; (dd) *Scolopodus filosus* Ethington & Clark, 1964, from Tu-37.4, NYSM 17437, ×155.

the *Rossodus manitouensis* Zone. As the uppermost Yudachica Member is not younger than the lower conodont Fauna B interval, the dendroids from the lower Río Salinas Member provide evidence that the lensing sandstone and conglomerate at the Yudachica–Río Salinas contact marks a biostratigraphically significant depositional sequence boundary at the Cambrian–Ordovician boundary.

5.b. Middle Río Salinas Member

A moderately diverse euconodont fauna (seven species) was recovered from limestones and calcareous sandstones in the middle Río Salinas Member at locality Tu (35.7–43.0 m; Table 2). This interval is readily correlated by five euconodont species that occur in the upper lower Tremadoc *Paltodus deltifer deltifer* Subzone in Baltica (see Löfgren, 1993, 1997; Löfgren, Repetski & Ethington, 1999) (Fig. 4). The association of the eponymous form *P. deltifer deltifer* (Lindström, 1955) (Fig. 8a–q) with *Paroistodus numarcuatus* (Lindström, 1955) (Fig. 8r–t), *Variabiloconus variabilus* (Lindström, 1955) (Fig. 8u–cc), and early representatives of *Drepanodus arcuatus* Pander, 1856 (Fig. 9h, i) characterizes *P. deltifer deltifer Subzone* faunas in these cool-water Baltic assemblages (Löfgren, 1993, 1997). *Variabiloconus transiapetus* Löfgren, Repetski & Ethington, 1999 (Fig. 9f, g) also occurs in the Swedish *P. deltifer deltifer* Subzone, and its occurrence in unrestricted marine successions in the western Great Basin has been used in correlating the subzone into the *Rossodus manitouensis* Zone in Laurentia (Löfgren, Repetski & Ethington, 1999). This correlation of the middle Río Salinas Member and Baltic *P. deltifer deltifer* Subzone with the *R. manitouensis* Zone is compatible with recovery of a single hyaline element of *Acanthodus uncinatus* Furnish, 1938 (Fig. 9v), a characteristic *R. manitouensis* Zone species in Laurentia (e.g. Landing, Westrop & Knox, 1996), but which is reported to range down into lowest Ordovician strata (Miller *et al.* 2003). The single element of *Scolopodus filosus* Ethington & Clark, 1964 (Fig. 8dd), from the Río Salinas Member represents a low occurrence of an uncommon species previously reported only as low as strata just above the *R. manitouensis* Zone in Laurentia (Ethington & Clark, 1981; Repetski, 1982). Numerous elements of *Cornuodus*? *clarkei* Landing sp. nov. (Figs 9j–u, 10r, s) were recovered from the *P. deltifer deltifer* Subzone, but the biostratigraphic significance of the form is currently undetermined as it does not seem to have been recovered from other successions.

6. Discussion

6.a. Conodont-based correlations

Despite the abundance of conodont elements in the Tiñu Formation, relatively few taxa are represented in each sample (Tables 1, 2). The conodonts and dendroid graptolites, however, show that the Tiñu Formation consists of an Upper Cambrian Yudachica Member and an unconformably overlying, upper lower Tremadoc Río Salinas Member.

Table 2. Upper lower Tremadoc conodonts from productive samples from the middle Río Salinas Member of the Tiñu Formation

	Tu -35.7	Tu -35.9	Tu -37.4	Tu -43.0
PROTOCONODONTS				
'*Phakelodus elongatus*'	29	2		1
Phakelodus tenuis	4			
PARACONODONTS				
Furnishina furnishi		11		2
Prooneotodus gallatani		5	1	6
PARACONODONTS?				
Coelocerodontus bicostatus				
asymmetrical tricostate el.		2		4
symmetrical (tetracostate) el.		2		
EUCONODONTS				
Acanthodus uncinnatus				1
Cornuodus? *clarkei* sp. nov.				
drepanodiform (symmetrical) el.		10	29	34
drepanodiform (subsymmetrical) el.			13	
scandodiform el.				5
paltodiform (costate) el.		2	7	15
acodiform-like el.				5
Drepanodus arcuatus				
drepanodiform (arcuatiform) el.	1	7	2	4
oistodiform (pipaform) el.			1	4
Paltodus deltifer deltifer				
oistodiform el.		3		12
scandodiform el.		1	5	13
suberectiform el.		3	1	1
drepanodiform (symmetrical) el.			4	2
drepanodiform (subsymmetrical) el.		2	17	
acodiform-like el.	1			25
Paroistodus numarcuatus				
oistodiform el.		1		
drepanodiform el.		4		
Scolopodus filosus			1	
Variabiloconus transiapetus				
scandodiform el.		3		
Variabiloconus variabilus				
drepanodiform (symmetrical) el.		16		
drepanodiform (asymmetrical) el.		5	1	
asymmetrical (laterally unicostate) el.		4		
acondiodiform el.		2		2

Acid-disaggregated samples had a 6.0 kg mass. el. – element.

One consequence of the low-diversity conodont faunas is that the finely subdivided, 'standard' conodont zonation and subzonation appropriate to the upper Upper Cambrian of tropical carbonate platform successions in western Laurentia (Miller, 1969, 1980; Miller *et al.* 2003) is not applicable to the Tiñu Formation. Thus, a 'simplified' conodont zonation with a coarser biostratigraphic resolution comparable to that appropriate in the cool-water Upper Cambrian of Baltica is appropriate to the Yudachica Member (Fig. 4).

A *Cordylodus andresi* Zone comparable to that recognized in the cool-water sequences of Baltica (Szaniawski & Bengtson, 1998) is appropriate for the

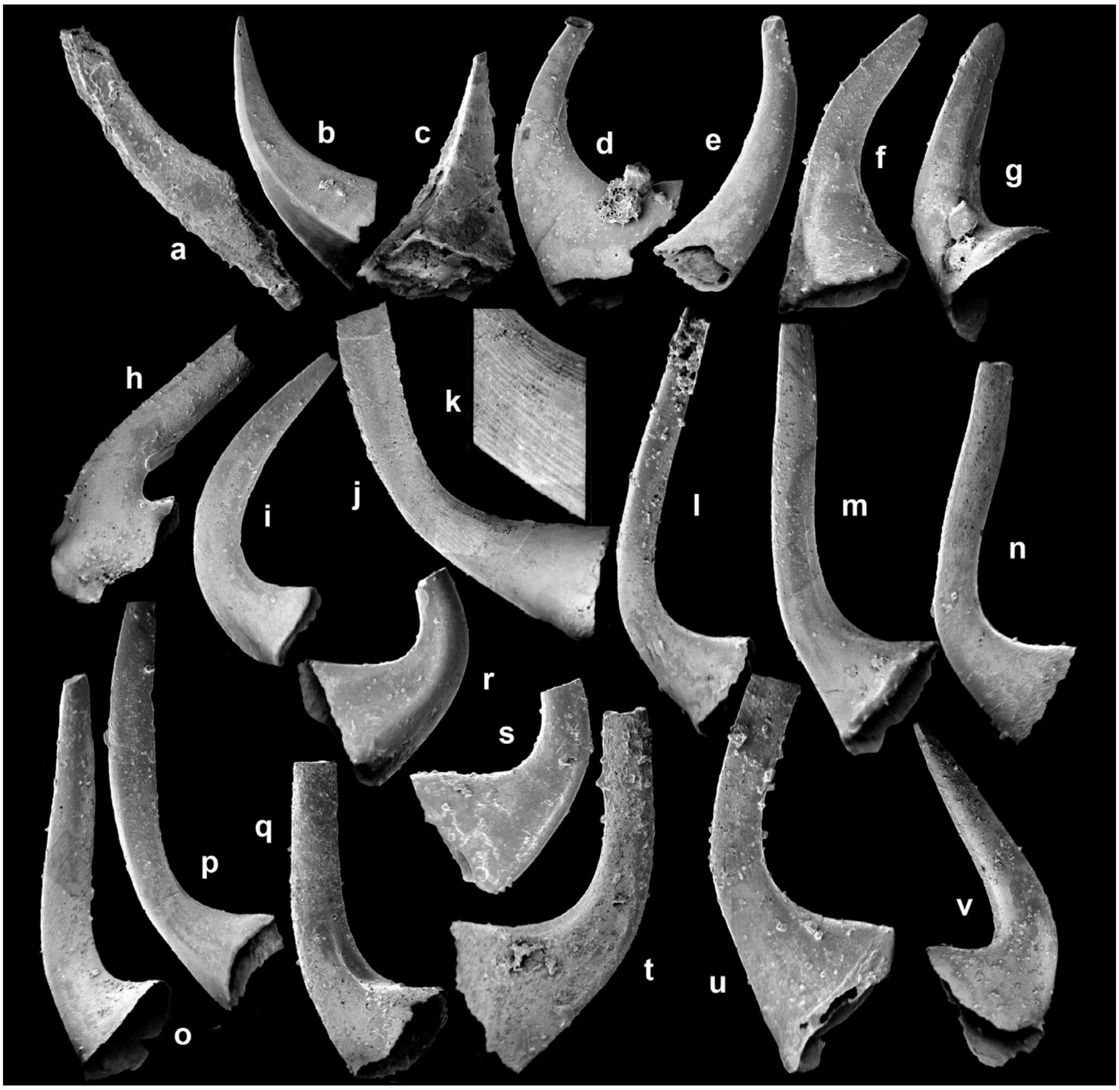

Figure 9. Lower Ordovician (upper lower Tremadocian) conodonts from the middle Río Salinas Member of the Tiñu Formation; hypotype specimens unless otherwise indicated. (a, b) *Coelocerodontus bicostatus* van Wamel, 1974. (a) Symmetrical tetracostate element, NYSM 17438, ×119, Tu-35.9; (b) asymmetrical (laterally unicostate) element, NYSM 17439, Tu-43.0, ×67. (c) *Furnishina furnishi* Müller, 1959, posterolateral view, NYSM 17440, Tu-43.0, ×88. (d, e) *Prooneotodus gallatini* (Müller, 1959) from Tu-35.9, NYSM 17441 and NYSM 17442, ×71 and ×90. (f, g) *Variabiloconus transiapetus* Löfgren *et al.* 1999, from Tu-35.9, proclined long- and reclined short-based scandodiform elements, NYSM 17443, ×91, and NYSM 17444, ×77. (h, i) *Drepanodus arcuatus* Pander, 1856, from Tu-43.0. (h) Oistodiform ('pipaform') element, NYSM 17445, ×68; (i) elongate drepanodiform ('arcuatiform') element, NYSM 17446, ×75. (j–u) *Cornuodus*? *clarkei* Landing sp. nov. from Tu-43.0 unless otherwise indicated. (j–l, n, u) Symmetrical drepanodiform elements. (j, k) Holotype NYSM 17447, lateral view, × 120 (see Fig. 10s), and enlarged view of longitudinal microstriae, ×280; (l) paratype NYSM 17448, ×99; (n, u) paratypes NYSM 17449 and NYSM 17465, ×100. (m, p, q, t) Subsymmetrical drepanodiform elements. (m) Paratype NYSM 17450 with weak posterolateral sulcus extending above zone of maximum curvature, ×120; (p) paratype NYSM 17551 with weak anterolateral sulcus and costa, ×98; (q) paratype NYSM 17452 with deep posterolateral sulcus, ×125; (t) paratype NYSM 17453 with weak posterolateral sulcus and very low anterolateral costa, ×111, Tu-37.4. (o) Scandodiform paratype NYSM 17454 with weakly laterally deflected base, ×95. (r, s) Acodiform-like elements. (r) Outer-lateral view showing rounded, medial costa, paratype NYSM 17455, ×79 (see Fig. 10r); (s) inner-lateral view showing anterolateral costa, paratype NYSM 17456, ×117. (v) *Acanthodus uncinatus* Furnish, 1938, subsymmetrical element with weak lateral deflection of anterior keel at aboral margin, NYSM 17457, ×88, Tu-43.0.

conodont fauna of the lower Yudachica Member. As the lowest *c.* 2.0 m of the Yudachica Member did not yield biostratigraphically useful conodonts, Upper Cambrian marine onlap across the Oaxacan Complex is tentatively referred to the *Cordylodus andresi* Chron (Fig. 4).

A few samples from the middle Yudachica Member have low-diversity conodont assemblages with the lowest *Cordylodus proavus* elements, and are assigned to the *C. proavus* Zone. A finer biostratigraphic resolution is not possible. The five subzones of the terminal Upper Cambrian *Cordylodus proavus* Zone as originally defined by Miller (1969, 1980) (Fig. 4) are based on species of *Clavohamulus* Furnish, 1938, *Fryxellodontus* Miller, 1969, and *Hirsutodontus* Miller,

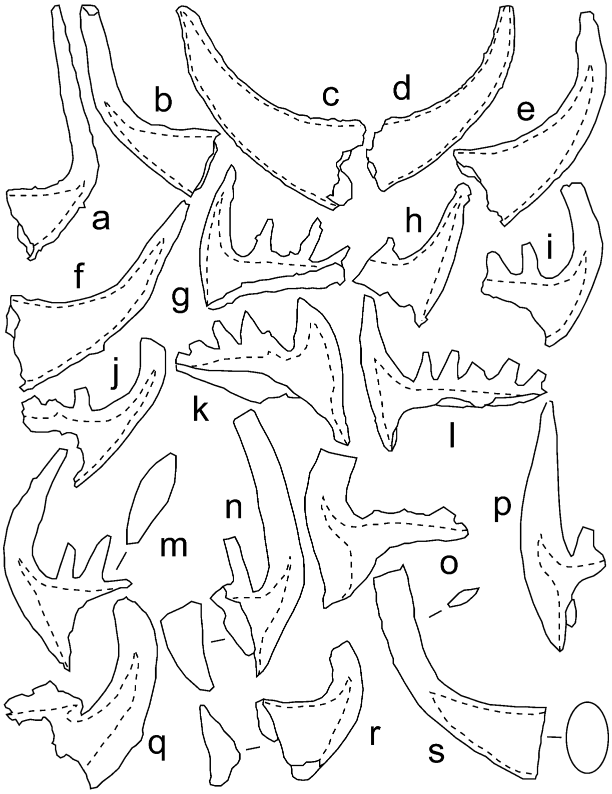

Figure 10. Conodont elements from the Tiñu Formation; outline of basal cavities dashed; hypotype specimens unless otherwise indicated. (a, b) *Orminskia rexroadae* Landing gen. et sp. nov. (a) Holotype NYSM 17390, ×54 (see Fig. 6a); (b) paratype NYSM 17393, ×80 (see Fig. 6e). (c) *Proconodontus serratus* Miller, 1969, NYSM 17407, ×52 (see Fig. 6t). (d) *Proconodontus muelleri* Miller, 1969, NYSM 17403, ×64 (see Fig. 6o). (e, f) *Eoconodontus notchpeakensis* (Miller, 1969); symmetrical (rounded) and asymmetrical (laterally deflected) elements, NYSM 17403 and 17404, ×60 and ×72, respectively (see Fig. 6q, r). (g, h) *Cordylodus andersi* Viira & Sergeeva *in* Kaljo *et al.* 1986; (g) rounded element, NYSM 17372, ×64 (see Fig. 5d); (h) compressed element, NYSM 17374, ×47, (see Fig. 5f). (i–l) *Cordylodus proavus* Müller, 1959. (i) Rounded element, NYSM 17376, ×49 (see Fig. 5h); (j) rounded element with secondary (‘*Cordylodus lindstromi*’) basal cavity tip, NYSM 17377, ×44 (see Fig. 5i); (k) compressed element with secondary (‘*Cordylodus lindstromi*’) basal cavity tip, NYSM 17379, ×35 (see Fig. 5k); (l) compressed element, NYSM 17380, ×42 (see Fig. 5l). (m–p) *Cordylodus caseyi* Druce & Jones, 1971. (m) Rounded element, NYSM 17381, ×40 (see Fig. 5m); (n) subsymmetrical rounded element, NYSM 17387, ×39 (see Fig. 5s); (o) rounded (‘*Cordylodus drucei*’) element, NYSM 17384, ×29 (see Fig. 5p); (p) compressed (cyrtoniodiform) element, NYSM 17382, ×31 (see Fig. 5n). (q) *Cordylodus prion* Lindström, 1955, *sensu* Nicoll (1991), rounded element, NYSM 17389, ×59 (see Fig. 5U). (r, s) *Cornuodus*? *clarkei* Landing sp. nov. (r) Paratype acodiform-like element, NYSM 17455, ×40 (see Fig. 5r); (s) holotype, NYSM 17447, ×60 (see Fig. 5j, k).

1969, which were not recovered from the Tiñu Formation. These genera and species are uncommon in unrestricted marine settings, such as the continental slope successions marginal to Laurentia (Landing, 1983; Landing, Barnes & Stevens, 1986) and unknown in the cool-water successions of Avalon and Baltica (Landing, Taylor & Erdtmann, 1978; Bruton, Koch & Repetski, 1988; Szaniawski & Bengtson, 1998). The presence of low-diversity conodont faunas that lack characteristic tropical/low-latitude uppermost Cambrian conodont genera and species is consistent with the apparent cool-water depositional setting of the Yudachica Member, and not with the low-latitude setting of Laurentia in the terminal Upper Cambrian.

The uppermost Yudachica Member is referred to the terminal *Cordylodus proavus*–lower Fauna B interval. The association of *Cordylodus caseyi* emend. and *Semiacontiodus nogamii* appears in the upper two subzones of the *C. proavus* Zone (Fig. 4), an interval which has been designated the *Cordylodus intermedius* Zone in many reports on the Laurentian and Chinese platforms (Bagnoli, Barnes & Stevens, 1987; Miller, 1984, 1988; Miller *et al.* 2003; Dong, Repetski & Bergström, 2004).

The upper Yudachica Member correlates into this so-called ‘*C. intermedius* Zone’, but we reject this biostratigraphic designation, and refer the lower part of the upper Yudachica to the upper *C. proavus* Zone for two important reasons (Fig. 4). The first reason is that *C. intermedius sensu formo* is apparently an element and junior synonym of the younger Tremadoc species *Cordylodus angulatus* (G. S. Nolan, unpub. Ph.D. thesis, Univ. Waterloo, Ontario, 1976; Nicoll, 1990; Landing, 1993). Thus, *C. intermedius* is inappropriate as an eponymous species for a terminal Cambrian interval. The second reason to suppress the so-called ‘*C. intermedius* Zone’ derives from the fundamental rules of stratigraphic nomenclature. Indeed, separating off the top two subzones of the *C. proavus* Zone as a ‘*C. intermedius* Zone’, while maintaining the name *C. proavus* Zone for the lower three subzones (Ethington & Clark, 1981; Bagnoli, Barnes & Stevens, 1987; Miller, 1988; Miller *et al.* 2003; Dong, Repetski & Bergström, 2004), creates a homonym, and is not acceptable (North American Stratigraphic Commission, 1983, article 19g). In short, the Tremadoc species *C. intermedius sensu formo* is not appropriate for a biostratigraphic zone that comprises the terminal Cambrian, while separating off the upper two subzones of the original *C. proavus* Zone (Miller, 1969, 1980) as a separate zone and retaining ‘*C. proavus*’ zone for the lower three subzones creates a homonym with the original definition of the *C. proavus* Zone.

6.b. Depositional history of Yudachica Member

Conodont biostratigraphy and lithofacies analysis of the Tiñu Formation help in a synthesis of the development of an Early Palaeozoic, passive margin succession on the Oaxacan Complex (Keppie *et al.* 2006). The

Yudachica Member marks latest Cambrian marine onlap in Oaxaca. It is significant that this thin member (12 and 13.3 m at SIx- and Tu) spans the lower–upper *C. andersi* Zone, a very thin, undifferentiated *Cordylodus proavus* Zone, and a terminal *C. proavus* Zone–lower Fauna B interval. The Yudachica is very condensed when correlated into coeval successions (upper *Proconodontus* Zone–lower Fauna B interval) in marginal Laurentian platform settings such as that of the western Great Basin (*c.* 150 m: Miller *et al.* 2003). Indeed, the Yudachica Member does not differ greatly in thickness from coeval sections in Baltica, and it is only 3 to 4 times as thick as the very condensed upper *Peltura* Zones–*Acerocare* Zone in the Oslo Region (e.g. Henningsmoen, 1957). U–Pb zircon geochronology shows that 4 Ma separate the middle *Peltura* Zones from the Cambrian–Ordovician boundary (Landing *et al.* 2000). Simple division of this figure into the thickness of the Yudachica Member yields an exceptionally low, average rate of accumulation (*c.* 0.3–0.4 cm ka^{-1}) of the compacted rock of the Yudachica Member.

The slow rate of development of accommodation space for Yudachica Member deposition may reflect either slow subsidence early in the cooling history of a rifted margin on the southern margin of the Rheic Ocean (e.g. Keppie & Ortega-Gutiérrez, 1999; Keppie *et al.* 2006) or deposition in the interior of a large, slowly subsiding cratonic block, perhaps following a strong eustatic rise. The resultant aggradation rate of the Yudachica Member seems to have been so low that relatively monofacial sequences persisted through the latest Cambrian at the southern and northern exposures of the Yudachica Member (storm wave-dominated to upper tidally influenced shelf at Tu and slope/ramp facies at SIx).

The low accumulation rate of the Yudachica Member and its apparent condensation may reflect unconformities and the presence of two or three depositional sequences in the member. An evaluation of possible depositional sequences within the Yudachica Member (Figs 3, 4) would require a more detailed, bed-by-bed analysis of the faunas and microfacies of the unit than was undertaken in this study. However, the presence of exceptionally thin conodont zones by comparison with those in other regions and the seeming coincidence of abrupt biostratigraphic changes (that is, appearance of new taxa and disappearance of taxa present in lower parts of the member) with apparent lithological breaks suggest a sequence stratigraphic architecture of the Yudachica Member.

Possible sequence boundaries may be marked at locality Tu by the appearance of a relatively thick, shale-dominated interval with the only large diagenetic (methanogenic?) nodules in the section (7.2–8.6 m) above the highest known *Cordylodus andresi* Zone fauna (Tu-7.2). This shale-dominated interval has limestones with an undifferentiated (lower?) *C. proavus* Zone fauna (sample Tu-8.5) near its top. Similarly, the appearance of a terminal *Cordylodus proavus* Zone fauna (sample Tu-9.0) is associated with the abrupt change into echinoderm tidalites in the upper Yudachica Member, and a sequence boundary may lie at the base of this facies.

Laterally equivalent horizons at locality SIx may include evidence for possible offlap and a sequence boundary marked by the thick, particularly coarse-grained debris flow (locally 7.1–7.35 m). This debris flow has an undifferentiated (lower?) *Cordylodus proavus* Zone fauna and lies above the highest known *C. andresi* Zone fauna (SIx-6.5). The proposed upper sequence boundary at Tu may be marked by a lenticular trilobite grainstone at SIx-8.0 with a terminal *C. proavus* Zone fauna. This latter possible sequence boundary lies at the base of a succession through the upper Yudachica Member that lacks the debris flows present in the lower part of the member.

These suggested depositional sequences in the Yudachica Member may correspond to several depositional sequences defined in the Ibex area on the western Laurentian margin (Miller *et al.* 2003). Interregional conodont-based correlations suggest correspondences between the Tiñu Formation *C. andresi* Zone and Ibex-area sequences 4 and 5, the undifferentiated (lower?) *C. proavus* Zone with sequences 7 and 8, and the terminal *C. proavus*–lower Fauna B interval with sequences 10 and lower 11.

Persistently low average aggradation rates of the Yudachica Member may help explain the faunal and lithological indications for low bottom-water oxygenation through most of the unit. Indeed, evidence exists for persistent dysoxia/anoxia of upper continental slope waters through the Late Cambrian and earliest Ordovician, and for the movement of these low-oxygen slope waters across platforms during times of particularly high eustatic sea levels (Landing, Benus & Whitney, 1992; Landing & Bartowski, 1996; Landing, Geyer & Bartowski, 2002).

In the late Late Cambrian (that is, *Saukia* Chron *sensu lato* in Laurentian successions), strong eustatic rise brought shorelines into the upper Mississippi River valley (e.g. Landing, Benus & Whitney, 1992). Thus, the simplest interpretation of the Yudachica Member is that it apparently reflects three particularly high sea-level intervals within the late Late Cambrian. Just as the highly condensed, organic-rich, black 'Alum Shale' facies records dysoxic/anoxic bottom environments in the interior of Baltica during Late Cambrian–earliest Ordovician high eustatic sea levels, most facies of the roughly coeval Yudachica Member, with exception of the echinoderm tidalites in the upper part of locality Tu, record bottom waters that were most likely mildly dysoxic by comparison with the highly dysoxic facies that appeared in coeval parts of the Alum Shale in Baltica (Schovsbo, 2001).

6.c. Depositional history of Río Salinas Member

A biostratigraphically resolvable depositional sequence boundary (lower Fauna B interval–lower *Rossodus manitouensis*) spans the Cambrian–Ordovician boundary at the Yudachica Member–Río Salinas Member contact. A thin conglomerate and abrupt facies change between the members mark this contact. With exception of minor (2–3 cm deep) erosional bevelling of the top limestone bed of the Yudachica Member, little evidence for stratigraphic cut-out at this sequence boundary is indicated at localities SIx and Tu. Robison & Pantoja-Alor (1968, fig. 3) measured a thicker section of the Yudachica Member at Arroyo Totoyac (16 m) than at localities SIx and Tu (11.95 m and 13.3 m, respectively: this report). These lateral changes in thickness may reflect differential erosion of the Yudachica Member before deposition of the Río Salinas Member. However, they did not report any change in trilobite faunas in the upper Yudachica member at Arroyo Totoyac, and it is possible that the locality only preserves a somewhat greater thickness of the lower conodont Fauna B interval.

Marine onlap and abrupt deepening is recorded by the olenid- and nematophorous dendroid-bearing, unburrowed, lower (but not lowest) Tremadocian dark-grey mudstones of the lower Río Salinas Member above its basal conglomerate. Investigation of this basal conglomerate was not undertaken, and it may prove to represent a deposit of a low-stand systems tract or a condensed transgressive systems tract.

This sea-level rise is apparently eustatic, as it correlates with sea-level rise within the *Rossodus manitouensis* Chron that drove shorelines into the upper Mississippi River valley region in Laurentia (Landing, Benus & Whitney, 1992; Ross & Ross, 1995; Landing, Westrop & van Aller Hernick, 2003; Landing & Westrop, 2006) and across Baltica during the ‘early to middle Tremadoc highstand interval’ (Nielsen, 2004). Development of a strongly dysoxic mid-water mass on continental slopes is recorded during this interval (middle–late part of the early Tremadoc) of high eustatic levels (Landing, Benus & Whitney, 1992; Landing, Westrop & van Aller Hernick, 2003), and an onlap of this low-oxygen water is apparently recorded by the lower Río Salinas Member.

A shoaling-up succession is recorded through the overlying siliciclastic mudstone-dominated, middle part of the Río Salinas Member. Improved bottom-water oxygenation is suggested by the colour change from dark-grey to grey-green mudstone, and is indicated by the appearance of intense burrow churning of the mudstone at about the level of the first limestone (Tu-35.7), and by the appearance of trilobite assemblages with asaphids and remopleurids in slightly higher limestone beds (Tu-43).

This facies change and the appearance of wave-generated fabrics are within the upper lower Tremadocian *Paltodus deltifer deltifer* Subzone. The near identity of conodont assemblages from the middle Tiñu Formation with *P. deltifer deltifer* Subzone assemblages from Poland and Sweden (Dzik, 1976; Löfgren, 1993, 1997; Löfgren, Repetski & Ethington, 1999) and their dissimilarity with coeval Laurentian faunas reflects both the degree of faunal exchange between these regions and their environmental similarities in the Early Ordovician. In lieu of any particularly useful palaeoclimatic evidence, the conodont faunas from the middle Río Salinas Member are most suggestive of temperate, unrestricted marine environments.

Shallow siliclastic shelf deposition is recorded through the higher strata of the Río Salinas Member, with interbedded mudstones and sandstones of a proximal tempestite facies capped by well-oxygenated, reddish purple shale and tidalite sandstones at the top of the section. However, we did not recover any biostratigraphically useful fossils above the *P. deltifer deltifer* Subzone limestones and sandstones. If no temporally significant stratigraphic discontinuities exist above this fossiliferous upper lower Tremadocian interval, the upper Tiñu Formation may provide a record of the highstand facies of the early–middle Tremadoc eustatic highstand and, possibly, the late Tremadoc–early Arenig eustatic lowstand (see Nielsen, 2004). In short, just as the Yudachica Member is interpreted to reflect a submergence of the Oaxaca Complex coeval with particularly high eustatic levels that brought shorelines into the upper Mississippi River valley, the Río Salinas Member reflects resubmergence of parts of Oaxaquia during the middle–late early Tremadoc eustatic high.

Unlike the Yudachica Member, no lithological or biostratigraphic evidence for exceptionally slow accumulation rates exists in the Río Salinas Member. A very rough comparative figure for the accumulation rate of the Río Salinas can be calculated. If the exposed, *c.* 60 m thick Río Salinas Member at locality Tu represents most of the 6 Ma long Tremadoc Epoch (Landing *et al.* 2000), its rate of accumulation of compacted rock is almost three times as fast as that of the Yudachica. In addition, the dominant facies (condensed interbedded limestones and siliciclastic mudstones of the Yudachica v. siliciclastic mudstones and sandstones of the Río Salinas) strongly differ between the two members, and there is an abrupt lithofacies contact between the members. These contrasts are interesting because both members were deposited during separate intervals of particularly high sea levels in the Late Cambrian and early Early Ordovician when Laurentian shorelines reached the upper Mississippi River valley. The difference between the two members in the lithofacies deposited and rate of production of accommodation space during times of comparable sea-level rise probably reflect epeirogenic changes on the Oaxacan passive margin that included increased availability of siliciclastic sediment and increased subsidence rates.

6.d. Palaeogeographic implications

Restudy of the Tiñu Formation, the lowest cover unit on the middle Proterozoic Oaxaca Complex, emphasizes that its Upper Cambrian–lowest Ordovician biotas and lithofacies do not support a palaeogeographic proximity to such tropical palaeocontinents as Laurentia (see Scotese & McKerrow, 1990; Sedlock, Ortega-Gutiérrez & Speed, 1993).

As noted by Shergold (1988), olenids are a significant proportion of the trilobites of the Tiñu Formation. Almost 40 % of the 21 genera reported by Robison & Pantoja-Alor (1968) are olenids, and this has influenced some biogeographic assessments of the fauna. For example, Shergold (1988) used the diversity of olenids to assign Oaxaca to the Baltic Province, which he interpreted as a set of cool-water biofacies. We agree that the trilobite faunas of Baltica, Avalonia and Gondwanan South America and Oaxaca likely lived under temperate conditions. However, Shergold's approach to trilobite biogeography focused on the biostratigraphic significance of faunas and does not lead to an improved understanding of global palaeogeography. Associated non-olenid taxa need to be considered, as well as the fact that olenid distribution reflects the wide occurrence of dysoxic water (e.g. Schovsbo, 2001) both at shallow depths at high latitudes and in deeper settings at mid- to low latitudes. Taylor (1977) attributed this wide latitudinal distribution of olenids to their preference for cold water. However, his approach assumed a 'modern ocean' with a source of cold polar water, a situation that likely did not exist in the Cambrian–Early Ordovician, an interval with no evidence for polar ice caps (Landing, Benus & Whitney, 1992). As he recognized, Shergold's (1988, p. 364) assessment actually reflected similarities in environmental conditions, and his palaeogeographic maps (Shergold, 1988, fig. 1) and those of Fortey & Cocks (2003, fig. 6) place the Mexican terranes in Gondwanan or peri-Gondwanan positions.

In their original study, Robison & Pantoja-Alor (1968) commented upon the close relationship with faunas from Argentina (e.g. Harrington & Leanza, 1957), which was considered to be part of a 'core area' of Gondwana by Fortey & Cocks (2003). The dikelocephalid *Mictosaukia globosa* (Robison & Pantoja-Alor) is a genus that supports a Gondwana or peri-Gondwana of Oaxaquia, as the genus occurs in Australia (Shergold, 1975, 1991), Afghanistan (Wolfart, 1970), south China (Peng, 1984), north China (Endo & Resser, 1937; Qian, 1986) and, possibly, Thailand (Shergold, 1975).

The distributions of *Onychopyge* Harrington, 1938 (represented in the Tiñu by *O. sculptura* Robison & Pantoja-Alor) and *Koldinioidia* Kobayashi, 1931 (represented in the Tiñu by *K. sulcata* Robison & Pantoja-Alor, 1968) are broadly congruent with that of *Mictosaukia*, with both occurring in Australia (Shergold, 1972, 1975, 1991; Jell, 1985), South China (Lu *et al.* 1984; Peng, 1984, 1990*b*) and North China (Endo & Resser, 1937). *Onychopyge* has also been recorded from Argentina (Harrington & Leanza, 1957; Zeballo & Tortello, 2005).

The asaphid *Asaphellus* Callaway, 1877, has been reported from several palaeocontinents and terrains, although there is considerable variation in such features as the size and position of the palpebral lobes in species assigned to the genus (e.g. Robison & Pantoja-Alor, 1968, pl. 98; Peng, 1990*a*, pl. 6, figs 3–7, 9, 10, 15; Shergold 1991, pl. 9, figs 9–14; Fortey & Owens, 2001, fig. 7a, b), and monophyly seems doubtful. As recognized by Shergold (1975, p. 33), the pre-Tremadocian asaphid *Golasaphus* Shergold, 1972, is closely comparable to species attributed to *Asaphellus* by Robison & Pantoja-Alor, and this may be where the affinities of the Mexican species lie (e.g. compare Shergold, 1972, pl. 10, fig. 1, with Robison & Pantoja-Alor, 1968, pl. 98, figs 12, 14). If correct, this provides additional support for a Gondwanan to peri-Gondwanan position for the Oaxaca.

This trilobite-based Gondwanan affinity agrees with middle Proterozoic lead isotopes (Ruiz *et al.* 1999; Cameron *et al.* 2004), thermal histories (Keppie & Ortega-Gutiérrez, 1999; Keppie *et al.* 2004), and a late Proterozoic–Early Cambrian rifting history (Keppie *et al.* 2006) that indicate a similarity of the Oaxacan Complex to the Santa Marta and Garzón massifs in the Columbian Andes (Keppie, 2004).

Upper Cambrian and lowest Ordovician conodont faunas from the Tiñu Formation are most similar to coeval faunas from Baltica and allow relatively precise biostratigraphic correlations into that palaeocontinent. However, as the olenid trilobites, this similarity reflects the ease of biotic exchange between temperate, unrestricted marine successions, and not palaeogeographic proximity of palaeocontinents in the Late Cambrian–Early Ordovician. It is important to note that conodonts from this time interval remain very poorly known from the temperate, unrestricted marine successions of Avalon (Landing, Taylor & Erdtmann, 1978). The recovery of a more complete Avalonian conodont record through this interval would likely document a succession of low-diversity conodont faunas with the same taxa as those known in the Tiñu Formation.

Lithofacies and faunas indicate movement of the West African margin of Gondwana from equatorial latitudes in the early Cambrian to the South Pole in the latest Ordovician (e.g. Landing, 1996, 2005). In the course of this southerly movement of West Gondwana, bedded carbonates became a limited component of platform successions as early Cambrian carbonate platforms disappeared, and bedded limestones were reduced to minor shell-hash beds. The Tiñu Formation provides the best-known record of these temperate, 'heterozoan' limestones in the Upper Cambrian and Lower Ordovician of West Gondwana. Similar

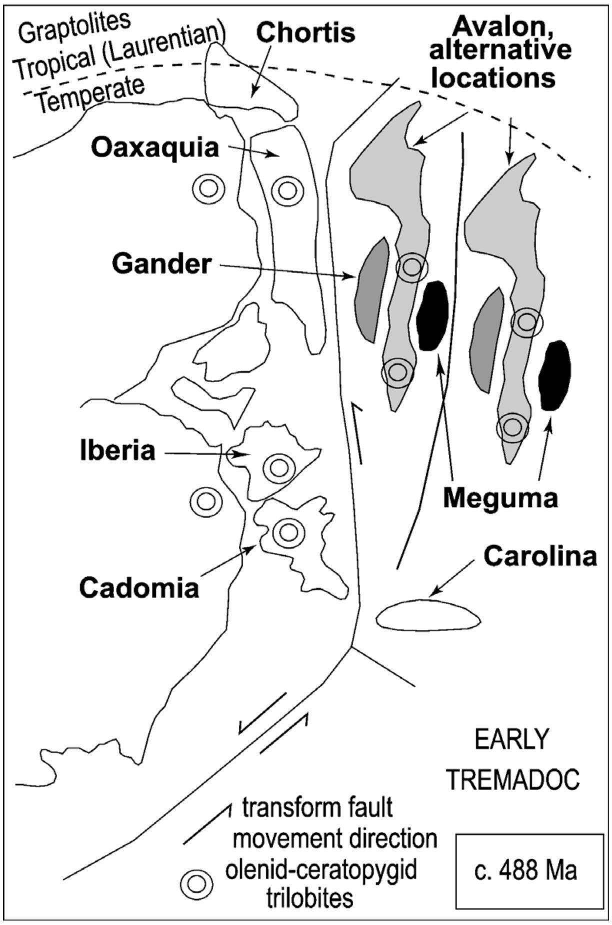

Figure 11. Generalized Cambrian–Ordovician boundary interval palaeogeography of southern hemisphere. Figure shows proximity of Oaxaquia to South American margin of West Gondwana. Iberia and Cadomia are contiguous with Moroccan margin of West Gondwana (Landing, 1996; Landing, Geyer & Heldmaier, 2006). After Avalon–Gondwana separation in latest Proterozoic times, increasing similarity of faunas and lithofacies of West Gondwana and Avalonia begins in the early–middle Cambrian boundary interval (*c*. 510 Ma) and is heightened into Early Ordovician times (e.g. Landing, 1996, 2005); this suggests that Avalon and West Gondwana came to lie at comparable palaeolatitudes, although the distance of their separation is undetermined.

Tremadoc limestones occur in a thick Cambrian–Lower Ordovician succession on the Garzón massif in Andean Columbia (Harrington & Kay, 1951; C. S. Bridger, unpub. Ph.D. thesis, Univ. National de Columbia, Bogota, 1982). As lithological and faunal data on this 'mainland' Gondwanan succession are limited, those from the Tiñu Formation provide the best palaeogeographic information on the Early Palaeozoic location of the Mexican–northern South American margin of Gondwana (Fig. 11).

7. Systematic palaeontology

Many taxa have been well discussed elsewhere or are represented by a limited number of specimens, and are not discussed below. Illustrated specimens are reposited in the New York State Museum (NYSM) Palaeontology Collection.

Class CONODONTA Eichenberg, 1930
Order PROTOCONODONTIDA Landing, 1995
Genus *Phakelodus* Miller, 1984

Type species. Oneotodus tenuis Müller, 1959, from the Upper Cambrian of Wyoming (? = *Proconodontus elongatus* Zhang *in* An *et al.* 1983, from the Upper Cambrian of north China).

Discussion. Two species of *Phakelodus* with similar, very elongate elements have been recognized in a number of reports. They are distinguished by the transverse cross-sections of their elements. *Phakelodus tenuis* (see synonymy in Müller & Hinz, 1991) elements have a rounded-oval cross-section. By comparison, *P. elongatus* (Zhang *in* An *et al.* 1983) (see synonymy in Müller & Hinz, 1991) elements have a teardrop-shaped cross-section with a sharply rounded posterior margin. However, these taxa may be synonymous as elements of both species regularly co-occur in samples and have comparable stratigraphic ranges (e.g. An *et al.* 1983; Müller & Hinz, 1991). Both *Phakelodus* 'species' also co-occur in the Tiñu Formation (Table 1). Because *P. elongatus* may prove to be a junior synonym of *P. tenuis*, its elements are distinguished as '*P. elongatus*' in this report (Fig. 6u, Table 1).

Order CONODONTOPHORIDA Eichenberg, 1930
Genus *Cornuodus* Fåhraeus, 1966, emend. Dzik, 1994

Type species. Cornuodus erectus Fåhraeus, 1966, from the Middle Ordovician of southeastern Sweden (= *Drepanodus longibasis* Lindström, 1955, from the Lower Ordovician of southeastern Sweden).

Cornuodus? *clarkei* Landing sp. nov
Figures 9j–u, 10r, s

Holotype. NYSM 17447 from sample Tu-43.0, upper lower Tremadoc *Paltodus deltifer deltifer* Subzone, middle Río Salinas Member of the Tiñu Formation, Oaxaca State, Mexico (Fig. 9j, k).

Paratypes. NYSM 17448–17452 and NYSM 17454–17456 from Tu-43.0 and NYSM 17453 from Tu-37.4.

Diagnosis. Probable *Cornuodus* species known from elongate, proclined to slightly reclined, anteriorly and posteriorly keeled, laterally compressed, coniform elements that form a weakly differentiated symmetry transition series (symmetrical drepanodiforms; weakly differentiated scandodiforms; asymmetrical paltodiforms with short, low lateral costa in zone of maximum curvature; and acodiform-like elements); cusps of many elements wider antero-posteriorly above zone of maximum curvature and lack posterolateral costae.

Etymology. For David L. Clark, who first examined conodonts from the Tiñu Formation (*in* Robison & Pantoja-Alor, 1968) and who was an important influence on EL.

Description. Coniform elements opaquely albid above basal cavity that extends into zone of maximum curvature (Fig. 10r, s). Cusp of some elements conspicuously broadens above zone of maximum curvature (Fig. 9j, p); bases narrow to broad in lateral view (Fig. 9j, u). Elements elongate; proclined to gently reclined (Fig. 9j, o); cusp lanceolate with sharp anterior and posterior keels; aboral margin circular (Fig. 9j) to strongly laterally compressed ellipse (Fig. 10s) to strongly flattened triangle with rounded outer-lateral costa (Fig. 9r, s). Development of fine longitudinal striae highly variable, may cover all surfaces of element (Fig. 9k) or be present only on posterolateral flanks in zone of maximum element curvature; microstriae become obsolescent in broad band anterior to aboral margin (Fig. 9j). Sharp anterior and posterior keels appear anterior to aboral margin and extend to tip; fine costae never present immediately lateral to posterior keel; drepanodiforms symmetrical (Fig. 9j, n) and transitional into scandodiforms with weakly laterally deflected base (Fig. 9m); posterior costa of scandodiforms may continue onto posterolateral surface of base as fine ridge (Fig. 9u). Paltodiform-like elements with low, short, anterolateral to posterolateral costa in zone of maximum curvature (Fig. 9p, q); in a few elements, a short, low anterolateral costa originates near the anterior keel in the zone of maximum element curvature, and curves slightly posteriorly and onto middle of cusp before becoming obsolescent. Acodiform-like elements asymmetrical with laterally deflected anterior keel (Fig. 9s) and broadly rounded medial costa on opposite side (Fig. 9r).

Discussion. The available specimens are morphologically simple elements of a weakly differentiated apparatus that is most similar to that of the stratigraphically long-ranging (Tremadoc–Ashgill) form *Cornuodus longibasis* (see Löfgren, 1999). Similarities between *C. longibasis* and *C.*? *clarkei* sp. nov. include element coloration and shape (albid, elongate, coniforms with short to long bases, lanceolate cusps, narrow to broad bases that are not strongly deflected laterally, and basal cavity extending to zone of maximum element curvature). The elements of both species have longitudinal microstriae that tend to be restricted to the posterior cusp margin (see Löfgren, 1999, p. 182), although those of *C.*? *clarkei* sp. nov. may extend along all surfaces of the elements. Löfgren's (1999) illustrations of *C. longibasis* elements shows that many of its elements have the posterolateral keel flanked by a low longitudinal costa immediately to the right and/or left. These costae do not occur in *C.*? *clarkei* sp. nov. elements. The apical widening of *C.*? *clarkei* sp. nov. elements is apparently not present in *C. longibasis*. Most elements of *C.*? *clarkei* sp. nov. recovered from the Tiñu Formation collections are drepanodiform-like, and correspond to Löfgren's (1999) S-series elements. The laterally costate acodiform-like elements (Fig. 9r, s) may be homologous to the asymmetrical P-elements in *C. longibasis*.

Cornuodus? *clarkei* sp. nov. is referable neither to *Acanthodus* Furnish emend. Landing, Westrop & Knox (1996), which has hyaline or weakly albid elements, nor to *Laurentoscandodus* Landing *in* Landing, Westrop & Knox (1996), a genus with nongeniculate oistodiforms. *Scalpellodus* Dzik, 1976, emend. Landing, Westrop & Knox (1996) is an uppermost Cambrian–lower Middle Ordovician genus also known from longitudinally microstriated, simple, albid, short- to long-based coniform elements arranged in a weakly to strongly differentiated symmetry transition series (see Löfgren, 1978; Landing, Westrop & Knox, 1996). However, *Scalpellodus* elements typically have lateral keels. The oldest known *Scalpellodus* species, the uppermost Cambrian–lowest Ordovician *S. utahensis* (Miller, 1969), is known from less elongate, proclined scandodiforms with antero- and posterolateral keels commonly bordered by a longitudinal sulcus (see *Paltodus utahensis* in Miller, 1969, pl. 63, figs 33–40). *Scalpellodus utahensis* has rare symmetrical, laterally compressed, 'unicostate elements' (Miller, 1980, p. 35) similar to the drepanodiforms of the new species.

Scalpellodus longipinnatus (Ji & Barnes, 1994), which ranges from the terminal Cambrian into upper lower Tremadoc-equivalent strata coeval with those yielding C.? *clarkei* sp. nov. (that is, the *Rossodus manitouensis* Zone), also has very elongate elements. However, *S. longipinnatus* had laterally compressed, anteriorly and posteriorly keeled scandodiforms comparable to those in *C.*? *clarkei* sp. nov., but also had *S. utahensis*-like scandodiforms in which the costae are bordered by a deep sulcus (Landing, Westrop & Knox, 1996). At present, *S. longipinnatus* is known from tropical carbonate platform successions in Laurentia, North China, South Korea and, probably, Australia (Landing, Westrop & Knox, 1996, pp. 675–6). Upper Lower–lower Middle Ordovician species of *Scalpellodus* include three Baltoscandian taxa: *S. viruensis* Löfgren, 1978, with similar but less elongate elements than *C.*? *clarkei* sp. nov.; *S. gracilis* (Sergeeva, 1964), in which the scandodiforms have very elongate bases and a strongly differentiated anterolateral costa; and *S. latus* (van Wamel, 1974), with drepanodiforms that are acontiodiform, not laterally compressed and lanceolate in cross-section as in *C.*? *clarkei* sp. nov. (see Löfgren, 1978).

Genus *Orminskia* Landing gen. nov

Type species. Orminskia rexroadae Landing gen. et sp. nov. from the Upper Cambrian (lower *Cordylodus andersi* Zone), lower Yudachica Member of the Tiñu Formation, Oaxaca State, Mexico.

Diagnosis. Euconodont genus with apparatus consisting of a weakly differentiated symmetry transition in hyaline, elongate, coniform elements with rounded cross-sections and deep basal cavities; elements lack surface micro-ornament in genotype species.

Etymology. For Edward A. Orminski (1924–2003), Anne Kaletta (*nee* Orminski) (1920–2005), Anna V. Orminski (1900–1992), and Helen B. Landing (*nee* Orminski) (1922–2006); maternal uncle, aunt, grandmother, and mother of EL, who encouraged an early interest in palaeontology.

Description. Apparatus consists of simple, elongate, proclined to reclined coniform elements (Fig. 6a–g) with circular to weakly laterally compressed cusps and bases, weakly differentiated symmetry transition series embraced by bilaterally symmetrical elements to elements with weakly laterally deflected bases (Fig. 6c); basal cavity deep and extends to zone of maximum cusp curvature (Fig. 10a, b); cusp completely lacks white matter and is transparent above tip of basal cavity; no surface micro-ornament present in elements of genotype species (Fig. 6d).

Comparison. Orminskia gen. nov. is known from exceptionally simple, coniform, euconodont elements. The form and basal cavities of its elements closely resemble those in *Teridontus nakamurai* (Nogami, 1967) (Fig. 6i, j, l). However, *Orminskia* gen. nov. elements have hyaline (glassy, transparent) cusps, and can be immediately distinguished from *T. nakamurai* elements with their opaquely white ('albid': see Müller, 1981, pp. W31–41) cusps in the same sample (Table 1). Scanning microscopy shows that the surfaces of *Orminskia* gen. nov. elements are smooth, while those of *T. nakamurai* have longitudinal microstriae on all external surfaces (Fig. 6k). Although a shallow longitudinal groove or subtle costa occurs on some *T. nakamurai* elements (e.g. Ji & Barnes, 1994, pl. 24, figs 4, 8, 9), these have not been noted on *Orminskia* gen. nov. elements. Thermal alteration and blackening of *Orminskia* new genus elements with deep burial or igneous intrusions (see Epstein, Epstein & Harris, 1977) would obscure their differences with *T. nakamurai* elements. It is likely that *Orminskia* elements have been recovered in earlier studies of Upper Cambrian rocks from orogenic belts, but have not been recognized.

Discussion. Orminskia gen. nov. occurs in the Upper Cambrian, and is the oldest known euconodont with hyaline elements. The absence or weak development of highly reflective albid matter in cusps and denticles is generally regarded as an important histological feature with significance in conodont taxonomy (e.g. Müller, 1981). Conodont genera with coniform to ramiform elements with hyaline or diffusely albid cusps are best known in the Ordovician from tropical carbonate successions (e.g. *Acanthodus* Furnish, 1938, emend. Landing, Westrop & Knox, 1996; *Eoneoprioniodus* Mound, 1965; *Multioistodus* Cullison, 1938; *Tropodus* Kennedy, 1980, *sensu* Bagnoli, Stouge & Tongiorgi, 1988; *Ulrichodina* Furnish, 1938, emend. Landing, Westrop & van Aller Hernick, 2003), but are also present in cool-water successions in the Baltic (e.g. *Scandodus* Lindström, 1955; *Scolopodus* Pander, 1856). These hyaline genera are presently assigned to a number of families seemingly arbitrarily assigned to superfamilies that include albid and hyaline taxa (see Clark *et al.* 1981, pp. W111–48). However, a shared histology of densely laminated apatite without internal spaces and pores may better reflect a phyletic and taxonomic relationship. If so, *Orminskia* new genus from the Upper Cambrian is potentially ancestral to Ordovician hyaline conodont genera.

Orminskia rexroadae Landing gen. et sp. nov
Figures 6a–h, 10a, b

Holotype. NYSM 17390 from sample Tu-2.05, Upper Cambrian (lower *Cordylodus andersi* Zone), lower Yudachica Member of the Tiñu Formation, Oaxaca State, Mexico (Fig. 6a).

Paratypes. NYSM 17391 (sample Tu-2.05), NYSM 17392 (Tu-4.3), NYSM 17393–17395 (Tu-4.95), and NYSM 17396, Upper Cambrian (lower *Cordylodus andresi* Zone), lower Yudachica Member of the Tiñu Formation, Oaxaca State, Mexico.

Diagnosis and description. As *Orminskia* gen. nov. is currently monospecific, the diagnosis of the type species and description of its elements correspond to that of the genus.

Etymology. For Carl B. Rexroad, who introduced EL to conodont studies during a National Science Foundation High School Summer Science Program in 1966.

Genus *Paltodus* Pander, 1856, emend. Lindström, 1971

Type species. Paltodus subaequalis Pander, 1856, from the Lower Ordovician of Estonia.

Paltodus deltifer deltifer (Lindström, 1955)
Figure 8a–q

1955 *Drepanodus deltifer* Lindström, p. 563, pl. 2, figs 42, 43.
1997 *Paltodus deltifer deltifer* (Lindström). Löfgren, pp. 264–5, figs 5Z–AG, 6H–N (source includes synonymy).

Material and occurrence. Hypotypes NYSM 17412–17426 from sample Tu-43.0, upper lower Tremadoc (*Paltodus deltifer deltifer* Subzone), middle Río Salinas Member of the Tiñu Formation, Oaxaca State, Mexico.

Discussion. The apparatuses of *Paltodus deltifer deltifer* and *P. deltifer pristinus* (Viira, 1970) have been illustrated and discussed by Szaniawski (1980) and Löfgren (1997) from Baltic successions. The abundant *Paltodus* elements from the middle Río Salinas Member have oistodiform elements with relatively elongate bases (Fig. 8a–e), suberectiforms with fairly prominent anterolateral costae (Fig. 8j, k), and acodiform-like elements with a lateral costa (Fig. 8q). These are features reported in *P. deltifer deltifer* (see Löfgren, 1997, fig. 6N, H, M, respectively). In addition, the distinctive scandodiform elements with broad, short cusps and antero-posteriorly elongated bases of *P. deltifer pristinus* (the 'varanguensiform' elements of Szaniawski (1980); see Löfgren (1997, fig. 5S)) were not recovered from the Tiñu.

Genus V*ariabiloconus* Landing, Barnes & Stevens, 1986

Type species. Paltodus bassleri Furnish, 1938, from the Lower Ordovician Oneota Dolostone of Minnesota.

V*ariabiloconus transiapeticus*
Löfgren, Repetski & Ethington, 1999
Figure 9f, g

1999 *Variabiloconus transiapeticus* Löfgren, Repetski & Ethington, pp. 166, 168, 170–1, pl. 3, figs 1–12, 14, 16–20, 22; text-fig. 3A–H.

Material and occurrence. Hypotypes NYSM 17443 and 17444 from Tu-35.9, upper lower Tremadoc (*Paltodus deltifer deltifer* Subzone), middle Río Salinas Member of the Tiñu Formation, Oaxaca State, Mexico.

Discussion. Single specimens of the proclined long- and reclined short-based scandodiform elements were recovered from the middle Río Salinas Member. Recovery of the species from the Tiñu Formation extends its range from the unrestricted marine successions of western Laurentia and Baltica (Löfgren *et al.* 1999) to the Mexican margin of Gondwana.

Variabiloconus variabilis (Lindström, 1955)
Figure 8u–cc

1955 *Oneotodus variabilis* Lindström, p. 582, pl. 2, figs 14–18, pl. 5, figs 4, 5; text-fig. 6.
1974 *Oneotodus variabilis* Lindström. Viira, 1974, p. 97, pl. 1, figs 14, 15; text-fig. 118.
1994 *Teridontus obesus* Ji & Barnes, p. 65, 66, pl. 24, figs 10–17; text-fig. 37B.
1999 *Variabiloconus variabilis* Lindström. Löfgren, Repetski & Ethington, p. 162, 164, 166, pl. 1, figs 1–20, pl. 2, figs 1–17; text-fig. 2A–S (source includes partial synonymy).

Material and occurrence. Hypotypes NYSM 17429–17432 and 17436 from Tu-37.4 and NYSM 17433–17435 from Tu-43.0, upper lower Tremadoc (*Paltodus deltifer deltifer* Subzone), middle Río Salinas Member of the Tiñu Formation, Oaxaca State, Mexico.

Discussion. A thorough description and illustration of the elements that compose the species' apparatus is in Löfgren *et al.* (1999). The elements from the Tiñu Formation are opaquely albid above the tip of a deep basal cavity that extends into the zone of maximum curvature. They are weakly to strongly reclined (Fig. 9u, x) and covered by longitudinal microstriae (Fig. 9w). A relatively rapid rate of tapering means the elements generally have a small cusp and large base (Fig. 9v–aa); however, a few elements are elongate (Fig. 9u). The drepanodiforms are symmetrical (Fig. 9u, v) or show slight a asymmetry with lateral deflection of the aboral margin (Fig. 9bb). The symmetrical acontiodiforms are not strongly antero-posteriorly compressed, and have weakly defined anterolateral costae (Fig. 9z, aa). Asymmetrical variants of the latter are laterally compressed, and have an anterolateral costa (Fig. 9cc) or a shallow posterior sulcus that is laterally displaced in the zone of maximum element curvature (Fig. 9y). The cross-section of most elements is circular to a laterally compressed oval, but sharply compressed cusps with narrowly rounded anterior and posterior margins of the distal end of cusp were recovered.

The elements of *Teridontus nakamurai* and *T. gracillimus* Nowlan, 1985, are more elongate, and include proclined variants (e.g. see Ji & Barnes, 1994). Lindström (1955) emphasized a rounded cross-section in his small collection of *Oneotodus variabilis* elements (40 specimens), but the larger collection from the Tiñu Formation includes forms with laterally compressed cusps. The rapidly tapering, reclined elements of the species, some of which have sharp anterior and posterior cusp margins, seem indistinguishable from those of *T. obesus*, and *T. obesus* is regarded as a junior synonym. '*Teridontus obesus*' occurs in somewhat older strata (*Cordylodus angulatus* Zone) in western Newfoundland (Ji & Barnes, 1994) than the middle Tiñu Formation. However, Löfgren (1997) and Löfgren, Repetski & Ethington (1999) recorded the species' lowest occurrence in the *C. angulatus* Zone in Sweden.

Class GRAPTOLITHINA Bronn, 1846
Order DENDROIDEA Nicholson, 1872
Family ANISOGRAPTIDAE Bulman, 1950
Genus *Rhabdinopora* Eichwald, 1855

Type species. Gorgonia flabelliformis Eichwald, 1840, from the Tremadocian of Estonia.

Rhabdinopora flabelliforme desmograptoides
(Hahn, 1912)
Figure 7

1912 *Dictyonema flabelliforme* var. *desmograptoides* Hahn, p. 139, pl. 20.
1954 *Dictyonema flabelliforme desmograptoides* Hahn. Bulman, pl. 5, figs 1–4; text-fig. 6a–c.

1987 *Callograptus* cf. *C. elegans* Hall. Sour & Buitrón, pp. 387–8, figs 1, 2.
1987 *Dictyonema flabelliforme* var. *oaxacensis* Sour & Buitrón, pp. 388–9, figs 3, 4.
1987 *Dictyonema flabelliforme* forma *typica* Brögger. Sour & Buitrón, pp. 389–90, fig. 5.
1987 *Dictyonema scitulum* Harris & Keble. Sour & Buitrón, p. 390, fig. 6.
1987 *Anisograptus delicatus* Cooper & Stewart. Sour & Buitrón, p. 391, fig. 7.
1987 *Bryograptus kjerulfi* Lapworth. Sour & Buitrón, pp. 391–2, fig. 8.
1987 *Staurograptus dichotomus* Emmons. Sour & Buitrón, p. 392, fig. 9.

Material and occurrence. Hypotypes NYSM 17408A, 17408B 17409, and 17411 from SIx-22.0 and NYSM 17410 from Tu-18.4, middle lower Tremadoc, lower Río Salinas Member of the Tiñu Formation, Oaxaca State, Mexico.

Description. Rhabdosomes relatively large, broken fragments reach 45 mm wide and 40 mm high, broadly conical with very broadly convex lateral margins (Fig. 7d, left side) and apparently parabolic apical end (Fig. 7a), length/breadth ratio about 1/1 in nearly complete specimens (Fig. 7a); 9–10 stipes/10 mm, with interstipe spaces wider than stipes (Fig. 7a, d); stipe branching primarily near apical end of rhabdosome (Fig. 7a, d); dissepiment spacing variable (6.5–7.5/10 mm but reaches 11.5/10 mm), regularly to irregularly spaced (Fig. 7b, c), wide spacing creates elongate interstipe spaces (Fig. 7a, d); locally ovoid mesh produced by thickening of dissepiments at junction with stipes (Fig. 7c, right side); autotheca number *c.* 15/10 mm (Fig. 7e, right three stipes).

Discussion. The dendroid specimens recovered from the lower Río Salinas Member at the Tu and SIx localities (Fig. 3) are referable to one morphologically defined subspecies of *Rhabdinopora*. Overall stipe and dissepiment spacing, relatively large rhabdosome size, the shape and variability of shape of interstipe spaces, and the lateral broadening of dissepiments match Bulman's (1954) definition of a subspecies known from Avalon (southern New Brunswick) and Baltica. All of these morphological features are evident in the six dendroid taxa that Sour & Buitrón (1987) recognized in a collection from the upper part of the exposed Río Salinas Member at locality SIx. Our dendroid collections from their locality (SIx-20, −20.5) and from a newly discovered horizon somewhat lower in the Río Salinas Member (Tu-18.4) suggest that Sour & Buitrón's (1987) taxa are best regarded as fragmentary taphonomic variants of one *Rhabdinipora* subspecies.

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Appendix 1. Members of the Tiñu Formation

Yudachica Member (new)

The Yudachica Member (new) comprises the lower Tiñu Formation. The Yudachica is a lithologically distinctive succession of thin- to medium-bedded limestone (rare mudstone and dominant fossiliferous wacke- to grainstones with intraclast granules and pebbles, with carbonate clast debris flows at Santiago Ixtaltepec) and light brownish green silt-shale with minor calcareous concretions. Thin calcareous quartz arenites are limited to the base of the member at its type section (Figs 1, 2). The Yudachica nonconformably overlies middle Proterozoic (*c.* 1 Ga) gneisses and granite at all localities; its top is an unconformity marked by a thin lensing sandstone and overlying limestone clast-dominated conglomerate bed or a phosphatized limestone clast conglomerate that forms the base of the overlying Río Salinas Member. The thickness of the Yudachica Member ranges from 12 m at Santiago Ixtaltepec (Fig. 1) to 16.4 m at Arroyo Totoyac (Pantoja-Alor, 1970, fig. 4). The Yudachica Member is named for the hamlet of Yudachica, approximately 0.5 km north of the type section of the Yudachica Member and Tiñu Formation at locality Tu (Figs 1, 2). ‘Yudachica’ has not been used previously for any Mexican stratigraphic unit. Fossils show that the Yudachica Member (new) is a condensed, upper Upper Cambrian succession interpreted to consist of three depositional sequences (Figs 2, 4).

Río Salinas Member (new)

The Río Salinas Member comprises the upper Tiñu Formation. This lithologically distinctive unit includes a thick, lower interval dominated by dark-grey and higher green-grey shale (55.4 m thick at the type section at locality Tu; Figs 1, 3). Isolated beds of medium- to thick-bedded, medium-grained sandstone and several trilobite-wacke- to packstone beds appear in the green-grey shale of the middle Río Salinas. The uppermost Río Salinas Member at locality Tu is composed of red silt-shale and trough cross-bedded quartz arenite with abundant orthoid brachiopods. The base of the member is an unconformity with the underlying Yudachica Member that is marked by a lensing, thin green quartz arenite (up to 10 cm thick) and an overlying conglomerate (up to 12 cm thick) composed of large, flat limestone and silt-shale clasts (locality Tu). This conglomerate is thin (7 cm) and phosphatized at locality SIx. The top of the Río Salinas is an unconformity locally overlain by Cretaceous limestone or Tertiary red beds at locality Tu and a thrust fault with lower Carboniferous sandstone at locality SIx. The thickest section (58.6 m) of the member is at the structurally undeformed type section of the Tiñu Formation. The top of the Río Salinas Member overlooks the deep Río Salinas valley to the north, and the member is named for this valley. 'Río Salinas' has not been used previously for any stratigraphic unit in Mexico. Fossils show that the lower–middle Río Salinas Member (new) is lower Lower Ordovician (middle–upper lower Tremadocian). Biostratigraphically useful fossils have not been recovered from the upper Río Salinas Member.