Geochemistry and U–Pb protolith ages of eclogitic rocks of the Asís Lithodeme, Piaxtla Suite, Acatlán Complex, southern Mexico: tectonothermal activity along the southern margin of the Rheic Ocean

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Abstract: Recent data indicating that the Piaxtla Suite (Acatlán Complex, southern Mexico) underwent eclogite-facies metamorphism and exhumation during the Devono-Carboniferous suggest an origin within the Rheic Ocean rather than the Iapetus Ocean. The Asís Lithodeme (Piaxtla Suite) consists of polydeformed metasediments and eclogitic amphibolites that are intruded by megacrystic granitoids. U–Pb (zircon) data indicate that the metasediments were deposited after c. 700 Ma and before intrusion of c. 470–420 Ma quartz-augen granite. The metasedimentary rocks contain abundant Mesoproterozoic detrital zircons (c. 1050–1250 Ma) and a few zircons in the range of c. 900–992 and c. 1330–1662 Ma. Their geochemical and Sm–Nd isotopic signature is typical of rift-related, passive margin sediments derived from an ancient cratonic source, which is interpreted to be the adjacent Mesoproterozoic Oaxacan Complex. Megacrystic granites were derived by partial melting of a c. 1 Ga crustal source, similar to the Oaxacan Complex. Amphibolitic layers exhibit a continental tholeiitic geochemistry, with a c. 0.8–1.1 Ga source (TDM age), and are inferred to have originated in a rift-related environment by melting of lithospheric mantle in the Oaxacan Complex. This rifiting may be related to the Early Ordovician drift of peri-Gondwanan terranes (e.g. Avalonia) from Gondwana and the origin of the Rheic Ocean.

The Acatlán Complex of southern Mexico (Fig. 1) contains eclogitic metasedimentary and metagneous rocks (Piaxtla Suite) that have been inferred to be vestiges of the Cambro-Ordovician Iapetus oceanic lithosphere (Ortega-Gutiérrez et al. 1999; Meza-Figueroa et al. 2003). This proposition was based on the interpretation that most of these rocks represent an ophiolitic complex that was subducted beneath Laurentia and underwent exhumation and dehydration to produce megacrystic granites (Esperanza granitoids dated at 440 ± 14 Ma: lower intercept U–Pb zircon age, Ortega-Gutiérrez et al. 1999: more recently dated at 471 ± 13 Ma: concordant U–Pb laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) zircon age, Sánchez-Zavala et al. 2004). Alternatively, Keppie & Ramos (1999) and Keppie (2004) suggested that the Acatlán Complex formed on the southern margin of the Ordovician–Carboniferous Rheic Ocean and did not collide with Laurentia until the Carboniferous during the amalgamation of Pangaea. To resolve this problem, which is fundamental to the Palaeozoic evolution of the Iapetus and Rheic oceans and to continental reconstructions, we have commenced a detailed field study of these eclogitic rocks combined with geochemistry and precise geochronology. In a companion paper, Middleton et al. (2005) have outlined the Carboniferous subduction and exhumation history of one of these eclogite-bearing units, the Asís Lithodeme; this is consistent with destruction of the Rheic Ocean but not with the Iapetus Ocean, which was closed by the Silurian. This paper presents geochemical and geochronological data for the Asís Lithodeme of the Piaxtla Suite, and focuses on the tectonic setting of the protoliths (Fig. 1). These data document a protolith association of continental tholeiites, continentally derived sediments and c. 470–420 Ma megacrystic granitoids inferred to form part of a rift–passive margin sequence, and that developed along the southern flank of the Rheic Ocean (northern margin of Gondwana: present coordinates) in Mexico (Oaxaquia terrane: Keppie 2004).

Geological setting

The Acatlán Complex of the Mixteca terrane in southern Mexico is juxtaposed on its eastern side against the c. 1 Ga Oaxacan terrane, and the terrane boundary is a Permian dextral flower structure (Fig. 1) (Elías-Herrera & Ortega-Gutiérrez 2002). Until recently, the Acatlán Complex was inferred to have undergone the following sequence of events (Fig. 2, left-hand side): (1) Cambro-Ordovician deposition of the clastic Petalcingo Group (Magdalena, Chazumba and Cosoltepec formations) and oceanic Piaxtla Group (Xayacatlán Formation and Esperanza granitoids); (2) Late Ordovician–Early Silurian polyphase
deformation during the Acatecan Orogeny, when the Piaxtla Group underwent eclogite-facies metamorphism and was thrust over the Petlalcingo Group metamorphosed at greenschist facies; (3) Devonian deposition of the Tecolote Formation; (4) Devonian deformation and greenschist-facies metamorphism during the Mixtecan Orogeny; (5) deposition of latest Devonian–Mid-Permian sedimentary rocks of the Otates and Patla-

Fig. 1. (a) Terrane map of Mexico showing the location of the Acatlán Complex (after Keppie 2004); (b) geological map of the Acatlán Complex (modified after Keppie et al. 2005) showing the location of the Asís area.

noaya formations (Ortega-Gutiérrez et al. 1999; Sánchez Zavala et al. 2000). Meza-Figueroa et al. (2003) have suggested that eclogitic rocks of the Piaxtla Suite at Piaxtla and Mimilulco lying on either side of the Asís area (Fig. 1b) represent mid-ocean ridge basalt (MORB), ocean island basalt (OIB) and island arc basaltic rocks that were metamorphosed to 11–15 kbar and 560 ± 60 °C (Fig. 1), conditions typical of low-to
medium-temperature eclogites developed in alpine-type tectonic settings (Carswell 1990).

However, recent work has shown the following (Fig. 2, right-hand side): (1) the Magdalena and Chazumba units are Permo-Triassic, whereas part of the Cosoltepec Formation is bracketed between rocks of c. 455 Ma age and the base of unconformably overlying uppermost Devonian sedimentary rocks (Vachard & Flores de Dios 2002; Derycke-Khatir et al. 2005; Keppie et al. 2006), whereas other parts of the Cosoltepec Formation are younger than c. 410 Ma (post-Silurian: Talavera-Mendoza et al. 2005); (2) the Tecomate Formation is of latest Carboniferous–Mid-Permian age (Keppie et al. 2004a); (3) the type-Xayacatlán igneous body is of earliest Silurian age and has a continental tholeiitic signature (Dostal et al. 2004); other units previously correlated with the Xayacatlán Formation are bracketed between c. 870 Ma and c. 470 Ma (Talavera-Mendoza et al. 2005); (4) the tectonothermal events are Devono-Mississippian (eclogite-facies metamorphism: Middleton et al. 2005), Permo-Triassic (greenschist-facies metamorphism: Malone et al. 2002; Keppie et al. 2004a, 2006), and Jurassic (local migmatization and high-temperature–low-pressure metamorphism: Keppie et al. 2004b); although Talavera-Mendoza et al. (2005) interpreted Ordovician granitoid ages in terms of Taconian–Salinian orogenic events, we suggest that they represent intrusive ages. This has led to reclassifying the Magdalena and Chazumba units as lithodomtes of a restricted Petalcingo Suite (Keppie et al. 2006). Similarly, the Piaxtla Group is replaced by the Piaxtla Suite containing the Ais Lithodeme, but excluding the megacrystic granitoids: the significant age difference between the Asis Lithodeme and the Xayacatlán unit suggests they should not be correlated.

**Lithologies and field relations**

The Ais Lithodeme of the Piaxtla Suite is composed mainly of metapsammitic and metapelitic rocks, and amphibolites, many of which are migmatized, which are intruded by megacrystic granites, the margins of which are mylonitized (Fig. 3). As all contacts are tectonic, the original relationships between the various rock types are uncertain. However, as the amphibolites

![Fig. 2. Time and space diagram showing the tectonostratigraphy of the Acatlán Complex after Ortega-Gutiérrez et al. (1999) and this paper. The Cosoltepec upper boundary allows for deposition after 410 Ma to accommodate recent data of Talavera-Mendoza et al. (2005).]

![Fig. 3. Geological map of the San Francisco de Asis area (18°27.6–29.4′N, 98°17.7–19.1′W).]
are always thin bands and lenses, they may have been either minor intrusive rocks or lavas within the metasedimentary rocks. The presence of xenoliths of the metasedimentary rocks within associated megacrystic granitoids could reflect an original intrusive relationship, but this remains to be confirmed. The occurrence of the quartz-augen granite within the K-feldspar megacrystic granite also suggests an original intrusive relationship that may, or may not, have been part of the same magmatic episode. The P–T paths of the different lithologies indicate that they share a similar retrograde path (Middleton et al. 2005).

The pelitic metasedimentary rocks consist mainly of quartz, plagioclase, biotite and phengite, minor garnet and rutile, and secondary chlorite, calcite and hematite. The metasomitic rocks are composed mainly of quartz with minor plagioclase, biotite and muscovite, and accessory opaque minerals. Migmatisation of these metasedimentary rocks is preferentially developed in the pelitic metasedimentary rocks. The amphibolites are composed mainly of albite and aligned amphibole with rare omphacite inclusions, minor epidote, phengite, biotite, garnet, and quartz, and accessory hematite, chlorite, titanite, ilmenite, fluorate, and calcite. Migmatic leucosomes are present in many of the amphibolites. The megacrystic granitoids are composed of quartz and albite, with minor phengite and secondary chlorite. Megacrusts are generally K-feldspar or quartz, with minor garnet. Quartz-augen granite occurs as small bodies within the K-feldspar granite and within the migmatitic metasedimentary rocks (Fig. 3). A few aplitic sheets cut the granitoids and have essentially the same mineralogy with the addition of secondary epidote.

Geochronology

Analytical methods

Two representative samples were collected for U–Pb isotopic analysis. A psammitic gneiss (CET-2; UTM 74.675, 38.902) was analysed using LA-ICP-MS at the Natural History Museum, London, and a quartz-augen granite from the megacrystic granite unit (CET-3; UTM 73.489, 39.515) was analysed by sensitive high-resolution ion microprobe (SHRIMP) at Stanford University (Fig. 3). The psammitic gneiss is composed mainly of quartz, plagioclase, biotite and muscovite, with accessory garnet and opaque minerals. The quartz-augen granite is composed of quartz and plagioclase (albite), with minor muscovite (phengite) and secondary chlorite. Analytical methods and LA-ICP-MS U–Pb analyses are available online at http://www.geolsoc.org.uk/SUP18237. A hard copy can be obtained from the Society Library.

Zircons were separated from psammitic sample CET-2 at the Complutense University of Madrid following conventional techniques. Details of the separation procedure have been given by Fernández-Suárez et al. (2002) and Jeffries et al. (2003). Analytical instrumentation, analytical protocol and methodology, data reduction, age calculation and common Pb correction followed those described by Fernández-Suárez et al. (2002), Jeffries et al. (2003) and Murphy et al. (2004).

Selected zircon grains from the quartz-augen granite sample (CET-3) were mounted in epoxy, polished, imaged in cathodoluminescence and photographed under reflected light, then gold-coated prior to analysis. Uranium–lead isotopes were analysed by SHRIMP-RG (SHRIMP-reverse geometry) at Stanford University following techniques described by Degraaf-Surpless et al. (2002). Errors in Table 1 are quoted at 1σ and ellipses on concordia diagrams are shown at the 68.3% confidence level.

Results

Seventy-two analyses, all representing one analysis per grain, were performed on zircons from psammitic sample CET-2. Of those, seven were rejected based on the presence of features such as discordance >10%, high common Pb detected in the U–Pb, Th–Pb, Pb–Pb isotope ratio plots, and/or elemental U–Pb fractionation or inconsistent behavior of U–Pb and Th–Pb ratios in the course of ablation (see Jeffries et al. 2003). Figure 4 shows concordia plots and a combined binned frequency and probability density distribution plot for the sample. Where the analyses overlap concordia with MSWD of concordance <2, we assign a U–Pb concordia age (see Ludwig 1998) as the best age estimate. Where analyses are normally discordant (i.e. they plot below concordia), we assign the 207Pb/206Pb age and error, as we are confident that any discordance is not a result of excess common Pb in the analysis or analytically induced problems such as laser-induced elemental fractionation (for details see Jeffries et al. 2003). Consequently, these ages will approximate the ‘correct’ age, assuming a zero-age Pb-loss event, and there is a small danger that a non-zero-age thermal event could result in these ages representing minimum ages. However, the amount of discordance within these zircons is minor (Fig. 4) and therefore this phenomenon is unlikely to affect any of the main conclusions reached regarding this dataset.

As shown in Figure 4 most analyses of psammitic sample CET-2 yielded Mesoproterozoic ages with a predominance of zircons in the age range c. 1050–1250 Ma (c. 75% of analyses). Five zircons yielded older Mesoproterozoic ages at c. 1330, 1332, 1486, 1540 and 1560 Ma, and one zircon yielded a late Palaeoproterozoic age of c. 1662 Ma. Nine zircons yielded early Neoproterozoic (Tonian) ages between c. 900 and 992 Ma. Finally, only one grain (Fig. 4c) yielded a younger Neoproterozoic (Cryogenian) age of 705 ± 8 Ma, the youngest grain dated in the sample.

The SHRIMP analyses of zircons from the quartz-augen granite (CET-3) yielded concordant ages ranging from 470 to

<table>
<thead>
<tr>
<th>Sample and spot no.</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>206Pb* (ppm)</th>
<th>206Pb/238U % Error1</th>
<th>207Pb/235U % Error1</th>
<th>208Pb/206Pb % Error1</th>
<th>208Pb/206Pb</th>
<th>Ages (Ma)</th>
<th>ρ1</th>
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<tr>
<td>CET3-1</td>
<td>292</td>
<td>185.2</td>
<td>17.5</td>
<td>0.06949 0.870 4.8700 4.800 0.05090 4.700 433 234 0.18</td>
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<tr>
<td>CET3-2</td>
<td>406</td>
<td>205.4</td>
<td>26.1</td>
<td>0.07470 0.500 0.59790 1.500 0.05796 1.400 464 528 0.33</td>
<td>0.33</td>
<td></td>
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<tr>
<td>CET3-3</td>
<td>295</td>
<td>137.4</td>
<td>17.4</td>
<td>0.06821 0.600 0.51100 4.900 0.05440 4.900 425 386 0.12</td>
<td>0.12</td>
<td></td>
<td></td>
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<tr>
<td>CET3-4</td>
<td>468</td>
<td>353.0</td>
<td>30.3</td>
<td>0.07520 0.450 0.59130 1.400 0.05702 1.300 467 492 0.33</td>
<td>0.33</td>
<td></td>
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<tr>
<td>CET3-5</td>
<td>319</td>
<td>137.5</td>
<td>20.2</td>
<td>0.07353 0.550 0.56400 2.300 0.05570 2.200 457 439 0.24</td>
<td>0.24</td>
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<tr>
<td>CET3-6</td>
<td>2166</td>
<td>671.0</td>
<td>199.0</td>
<td>0.10602 0.290 1.02500 2.000 0.07010 2.000 650 931 0.15</td>
<td>0.15</td>
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206Pb* indicates radiogenic Pb concentrations.
1Errors are 1σ. Error in standard calibration was 0.34–0.55% (not included in above errors but required when comparing data from different sources).

References

Jeffries et al. (2003) Analytical instrumentation, analytical protocol and methodology, data reduction, age calculation and common Pb correction followed those described by Fernández-Suárez et al. (2002), Jeffries et al. (2003) and Murphy et al. (2004).

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420 Ma and two discordant analyses with $^{207}\text{Pb}^{206}\text{Pb}$ ages ranging from c. 234 Ma to c. 931 Ma (Fig. 5, Table 1).

**Interpretation**

The youngest c. 705 Ma detrital zircon from the metapsammite (CET-2) provides an older limit for the time of deposition. A younger limit of c. 470–420 Ma is provided by the presence of metasedimentary xenoliths in the quartz-augen granite, assuming that the xenoliths were derived from the country rocks.

Beyond broadly constraining the intrusive age of the mega-

**Geochemistry**

**Analytical methods**

Twenty samples of amphibolite, four samples of pelitic and psammitic gneiss, 18 samples of granite (seven megacrystic, 11 mylonitic), and 27 samples of migmatite were collected for chemical analyses. The samples were analysed by X-ray fluorescence spectrometry for major and several trace elements (Rb, Sr, Ba, Zr, Nb, Y, Zn, V, Cr and Ni) in the Nova Scotia Regional Geochemical Centre at Saint Mary’s University, Halifax. From this set, 24 samples (five amphibolite, two pelitic–psammitic gneiss, seven granite, 10 migmatite) were selected for additional trace elements analyses (rare earth elements (REE), Hf, Zr, Nb, Ta and Th) by ICP-MS in the Geochemical Laboratory of the Ontario Geological Survey in Sudbury, Ontario, and for Sm–Nd isotopic analyses, which were performed at Memorial University, Newfoundland. All geochemical and isotopic data are available online in supplementary files (see p. 686).

Precision and accuracy of the X-ray data have been reported by Dostal et al. (1995). To facilitate comparison between the various lithological units, $\varepsilon_{\text{Nd}}(t)$ values quoted in the text are for $t = 475$ Ma.

The use of chemical data of rocks metamorphosed under high-grade conditions for tectonic models assumes that the elements under consideration remained essentially immobile during secondary processes. There are several lines of evidence to suggest that most major elements, such as Si, Al, Mg, Fe and Ti, as well as many trace elements including high field strength elements (HFSE; Zr, Hf, Nb, Y, Ta, and Th), REE and
the transition elements (Cr, Ni and V) were not redistributed in the analysed rocks. In particular, their consistent trends and similarities to modern volcanic and sedimentary rocks suggest that they essentially retained the original characteristics.

Results

Amphibolites. With the exception of one sample (101-4, which shows evidence of significant crustal contamination), the amphibolites display remarkably uniform geochemistry and trends consistent with igneous processes. They have characteristics of differentiated tholeitic basalts, with SiO₂ between about 48 wt% and 52 wt%, FeO₉/MgO ranging from 1.5 to 3.5, high content of FeO (11.1–17.1 wt%), and strong correlations between FeO/ MgO, TiO₂, P₂O₅, V and Zr, indicating enrichment in these elements during fractionation (e.g. Figs 6–8) and Nb/Y between 0.2 and 0.75. The REE patterns (Fig. 9a) show a slight enrichment in light REE (LREE) with (La/Yb)ₙ ranging between 1.5 and 2.1, and (La/Sm)ₙ between 1.2 and 1.6 (sample 101-4 has higher LREE enrichment; La/Ybₙ = 8.7, La/Smₙ = 3.5). The shape of trace element patterns is more similar to enriched (E-) than normal (N-) MORB (Fig. 10a and b): N-MORB-normalized patterns are enriched in strongly incompatible elements, such as Th and LREE relative to heavy REE (HREE), whereas E-MORB-normalized patterns are relatively flat. These overall trends are typical of differentiated tholeitic basalts derived from an E-MORB source. In comparison with calc-alkaline and island arc mafic rocks, the amphibolites have wide ranges in Cr (34–328 ppm) and Ni (35–172 ppm), and also high Ti/V (35–45) and Ti/Zr (60–120), as well as relatively constant Zr/Y (4.0–6) values, with the exception of two samples with high Zr that have Zr/Y of 8.8 and 10.5. According to Pearce & Norry (1979), mafic rocks with Zr/Y >4 are typical of within-

![Fig. 6. SiO₂ v. MgO, FeO, and FeO/MgO for the amphibolites (a–c), and SiO₂ v. MgO, TiO₂ and FeO/MgO for the migmatitic rocks (d–f) of the Asis Lithodeme.](image)

![Fig. 7. (a) FeO/MgO v. TiO₂ (wt%), (b) FeO/MgO v. Zr and (e) V v. TiO₂ for amphibolites of the Asis Lithodeme The lines showing calc-alkaline and tholeiitic trends are after Miyashiro (1974).](image)

![Fig. 8. (a) FeO v. FeO/MgO, (b) FeO/MgO v. TiO₂ and (c) V v. TiO₂ for migmatitic rocks of the Asis Lithodeme.](image)

plate basalts (e.g. Fig. 11).

The Sm–Nd isotopic signature of three of four amphibolite samples is very similar, with εNd(t) ranging from +2.8 to +4.6 (t = 475 Ma), 1⁴⁷Sm/¹⁴⁴Nd from 0.16 to 0.18 and TDM from 1.09 to 1.27 Ga (Fig. 12a). These εNd(t) values are considerably lower than values expected for juvenile magmas from a depleted mantle source. The lack of a discernible negative Nb anomaly on the spidergram plots and low ratios, such as La/Nb (0.7–0.9) and Th/Ta (1.5–2.1), indicate that these low values are not due to crustal contamination and they are interpreted to reflect the original signature of the magma. The contamination of sample 101-4 is clearly indicated by its very low εNd(t) of ~8.3, low ¹⁴⁷Sm/¹⁴⁴Nd (0.12), high TDM (1.68 Ga), and high La/Nb (2.3).

The migmatized amphibolites are very similar to the amphibolites, having SiO₂ <55%, high contents of FeO (12.1–17.4 wt%), high FeO/MgO (1.6–2.8), strong correlations between FeO, TiO₂, P₂O₅, V and Zr (Fig. 8), and Nb/Y between 0.3 and 0.9, characteristics typical of differentiated tholeiites. Compared with the amphibolites, these mafic migmatites have similar wide ranges in Cr (82–311 ppm) and Ni (45–98 ppm), high Ti/V (35–
50) and Ti/Zr (80–120) as well as relatively high Zr/Y (4.0–10). The REE patterns also show (Fig. 9b) a slight enrichment in LREE, with (La/Yb)$_n$ ranging between 1.8 and 4.7 and (La/Sm)$_n$ between 1.2 and 2.1, and, as for the amphibolites, the shape of trace element patterns is more similar to E-type than N-type MORB (Fig. 10c and d). The Sm–Nd isotopic signature is also very similar, with $\varepsilon_{\text{Nd}}$ ranging from +2.0 to +5.8 ($t = 4.075$ Ma), $^{143}\text{Sm}/^{144}\text{Nd}$ from 0.15 to 0.18 and $T_{DM}$ from 0.75 to 1.25 Ga (Fig. 12). The lack of a discernible negative Nb anomaly on the spidergram plots and low ratios such as La/Nb (0.7–0.75) indicate that crustal contamination was insignificant, and so, as for the amphibolites, these values are interpreted to reflect an original signature that is typical of differentiated tholeiitic basalts derived from an enriched MORB source.

**Pelitic and psammitic gneiss.** The pelitic and psammitic gneiss is highly variable in chemistry, with SiO$_2$ ranging from 59.1 to 92.0 wt%. Two samples were analysed for REE and Sm–Nd isotopes, one with low SiO$_2$ (59 wt%) and one with high SiO$_2$ (86.5 wt%). These samples show similar chondrite-normalized patterns, with moderately enriched LREE (La/Sm$_n$ 3.1–4.1), and flat HREE (Dy/Yb$_n$ 0.8–1.2), although the more siliceous sample
has lower total REE and low La relative to Ce (Fig. 9c). Similar patterns are also exhibited on a MORB-normalized plot, with the exception of a strong positive Zr anomaly occurring in the more siliceous sample (Fig. 13a). Despite variations in geochemistry, the Sm–Nd isotopic signature of the two samples is very similar, with $\varepsilon_{\text{Nd}}$ values ranging from $-7.2$ to $-7.5$ at $t = 475$ Ma, $^{147}\text{Sm}/^{144}\text{Nd}$ from $0.12$ to $0.13$, and $T_{\text{DM}}$ from $1.68$ to $1.86$ Ga (Fig. 12).

The migmatized metasedimentary rocks have very similar characteristics to the pelitic and psammitic gneisses (Fig. 6d-f) with SiO$_2$ contents varying from 55 to 70 wt% SiO$_2$. The REE patterns are enriched in LREE ($\text{La}/\text{Sm}$) $n$ 3.8 to 4.0 with relatively flat HREE with ($\text{La}/\text{Yb}$) $n$ 1.9 to 8.1 (Figs 9d and e) and strong negative Eu anomalies. MORB-normalized trace element profiles are enriched in strongly incompatible trace elements, including Th and Ba, and LREE, relative to the HREE and HFSE. They display distinct negative anomalies for Nb and Ti (Fig. 13b and c), and have La/Nb values (1.1 to 1.8) typical of crustal input. $\varepsilon_{\text{Nd}}$ values are more depleted than the granitoid rocks, ranging from $-5.9$ to $-8.5$ at $t = 475$ Ma, with $^{147}\text{Sm}/^{144}\text{Nd}$ from $0.11$ to $0.12$ and $T_{\text{DM}}$ from $1.45$ to $1.73$ Ga. These are values that are typical of Mesoproterozoic continental crust. These $\varepsilon_{\text{Nd}}$ values are considerably lower than those of the amphibolites, the mafic component of the migmatitic amphibolites and the granitoid rocks, and are typical of an intra-crustal origin.

Although volumetrically minor, migmatites with SiO$_2$ content $>70$ wt% can be distinguished from less siliceous migmatites and from granitoid rocks by lower Fe$_2$O$_3$, MgO, CaO, TiO$_2$, Ni, V and Cr and by highly variable REE and HFSE abundances that are either similar to or lower than the 55–70 wt% pelitic

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**Fig. 11.** Zr/Y v. Zr discrimination diagram (after Pearce & Norry 1979) for the Ası´s Lithodeme amphibolites. IAB, island arc basalt; MORB, mid-ocean ridge basalt; WPB, within-plate basalt.

**Fig. 12.** Sm–Nd isotopic data for the Ası´s Lithodeme: Ası´s amphibolites, granitoid rocks and pelitic and psammitic gneisses; and the migmatitic rocks with typical Sm–Nd isotopic compositions of peri-Rodinian oceanic lithosphere (Murphy et al. 2000, 2004) and Mesoproterozoic continental crust (after Dickin & McNutt 1989, 1990; Patchett & Ruiz 1987; Daly & McLelland 1991; McLelland et al. 1993; Dickin 2000). (a) $\varepsilon_{\text{Nd}}$ v. time comparing Sm–Nd isotopic data; (b) $\varepsilon_{\text{Nd}}$ v. $^{147}\text{Sm}/^{144}\text{Nd}$ (at $t = 475$ Ma), comparing the Sm–Nd isotopic data with typical Sm–Nd isotopic compositions of Avalonian crust (Murphy & Macdonald 1993; Murphy et al. 1996) and the average upper crust, which is bracketed between modern global average river sediment ($^{147}\text{Sm}/^{144}\text{Nd} = 0.114$; $T_{\text{DM}} = 1.52$ Ga; Goldstein & Jacobsen 1988) and the average age of sedimentary mass (Miller et al. 1986). Iapetan crust includes normal and depleted island arc tholeiites, and ophiolitic complexes in Newfoundland and Norway (Pedersen & Dunning 1997; MacLachlan & Dunning 1998). Tińu Formation data are from Murphy et al. (2005). CHUR, Chondritic Uniform Reservoir.
migmatites (Figs 9e and 13c). Some samples display strong enrichment in Zr. However, they have similar εNd(t) and TDM values to migmatites with 55–70 wt% SiO2, although one sample (102-3), has anomalously high 147Sm/144Nd (Fig. 12). In general, the trace element, REE and Sm–Nd isotopic characteristics of the migmatites with >55% SiO2 are similar to those of pelitic and psammitic gneisses.

Granitoid rocks. The megacrystic and mylonitic granitic rocks display very similar geochemical signatures, with SiO2 ranging from 60.5 and 74.5 wt% and a decrease in FeOt, TiO2 and MgO with increasing SiO2 (Fig. 14). Chondrite-normalized REE patterns show moderate enrichment in LREE ((La/Yb)n of 2.5–5.32), relatively flat HREE with (La/Yb)n of 1.1–4.1, and exhibit strong negative Eu anomalies (Fig. 9f). MORB-normalized trace element profiles (Fig. 13d) are distinctly enriched in strongly incompatible trace elements, including Th and Ba, and are moderately enriched in LREE relative to the HREE and HFSE, and display distinct Nb and Ti negative anomalies. The geochemical characteristics of the dated sample, CET-3, are typical of the megacrystic and mylonitic rocks (e.g. Figs 9f, 12 and 13d). Although these rocks straddle the arc + syncollisional and within-plate boundary on the Nb–Y plot (Fig. 15a), they plot mainly in the arc field on the Rb–(Nb + Y) and Ta–Yb discrimination plots (after Pearce et al. 1984; Fig. 15b and c). The Sm–Nd isotopic signature is very different from that of the amphibolites, and is remarkably uniform, with εNd(t) ranging from −4.3 to −5.5 (t = 475Ma), 147Sm/144Nd from 0.12 to 0.14, and TDM from 1.5 to 1.8 Ga. Sample CET-3 lies within this range (Fig. 12). These εNd(t) values are considerably lower than those of the amphibolites and are typical of an intra-crustal origin rather than derivation by fractionation of a mafic parent (Fig. 12). A crustal origin is supported by the Nb and Ti anomalies on MORB-normalized plots, the low 147Sm/144Nd and high TDM. However, these values are notably higher than those for the pelitic and psammitic gneisses, suggesting that they are not simply derived from partial melting of host rocks, although the possibility of a mixed mafic–metasedimentary source cannot be excluded. Nevertheless, as the granitoid rocks are predominantly crustally derived, the geochemical data alone cannot distinguish between an origin in a coeval arc and inheritance from older crust that was itself formed in an arc environment.

Discussion

Metasedimentary rocks

The pelitic and psammitic gneisses and their migmatized equivalents show remarkable chemical similarities to the unmetamorphosed, Tremadoc Tinu Formation, which occurs to the east of the Acatlán Complex, where it rests unconformably upon the c. 1 Ga Oaxacan Complex (Fig. 1a). The Tinu Formation contains a Gondwanan fauna (Robison & Pantoja-Alor 1968) and is unconformably overlain by Carboniferous sedimentary rocks containing a Laurentian fauna (Boucot et al. 1997). The geochemistry of sandstones and shales of the Tinu Formation (Murphy et al. 2005) also shows a wide range in SiO2 that, as for the Ais high-grade metasedimentary rocks, is interpreted to reflect derivation from crust with mafic and felsic components. In both cases, the presence of a significant mafic component suggests proximal sources (e.g. Nesbitt & Young 1996). The Tinu sedimentary rocks display LREE enrichment with a moderate Eu anomaly (Fig. 9c), similarly low εNd(t) values (−7.1 to −7.7, calculated for t = 475Ma) and Mesoproterozoic TDM ages (1.5–1.83 Ga), and contain detrital zircons that yield ages of 990–1200 Ma (Gillis et al. 2001), i.e. similar to those of the Ais psammitic sample, CET-2. The Tinu Formation data are interpreted to reflect derivation mainly from the underlying Oaxacan Complex with little or no input from distal or juvenile sources. Minor input from the Amazon craton may be indicated by the c. 1300–1660 Ma detrital zircons, and the youngest c. 700 Ma detrital zircon could have come from either Amazonia or Avalonia (Fig. 16). The similarities in geochemistry, Sm–Nd isotopes and ages of detrital zircons between the high-grade metasedimentary rocks of the Ais Lithodeme and the Tinu Formation suggests that protoliths of the Ais metasedimentary
rifting-related and/or passive margin sedimentary rocks. Although poorly constrained, the c. 700–470 Ma age bracket for deposition of the Asís Lithodeme, together with indistinguishable geochemical, isotopic and detrital zircon population characteristics, is consistent with correlation of the protoliths with the Tremadoc Tinu Formation, and therefore, as for the Tinu Formation, an origin along the Gondwanan margin of the Rheic Ocean.

Amphibolites

The amphibolites and migmatitic amphibolites have geochemical signatures typical of differentiated continental tholeiites, with relatively little evidence for crustal contamination suggesting a thinned crust and/or an extensional setting. Elias-Herrera et al. (2004) reported ages of 442 ± 2 Ma for a zircon core with rims dated at c. 360–345 Ma (U–Pb SHRIMP ages) that are here interpreted as protolith and eclogite-facies metamorphism, respectively. The latter age is similar to c. 346 Ma age of metamorphism reported by Middleton et al. (2005), and the c. 442 Ma age falls within the range of ages determined for the quartz-augen granite (this paper) and other megacrycstic granitoids in the Acatlán Complex (Sánchez-Zavala et al. 2004; Miller et al. 2005). This suggests that the amphibolites and the granitoids in the Asís Lithodeme are coeval and represent a bimodal suite.

The $T_{DM}$ ages of the amphibolites range from 0.75 to 1.27 Ga, indicative of derivation from continental mantle lithosphere. Such $T_{DM}$ ages are common in igneous complexes along the northern Gondwanan margin, and are interpreted to reflect ancestral crust and mantle lithospheric components that formed in the peri-Rodinian ocean between 1.2 and 0.75 Ga, were accreted to the Amazonian margin of Gondwana in the late

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**Fig. 15.** (a) Nb vs Y, (b) Rb vs Nb + Y and (c) Ta vs Yb discrimination plots for granitoid rocks of the Asís Lithodeme (after Pearce et al. 1984). WPG, within-plate granite; VAG, volcanic arc granite; syn-COLG, syncollisional granite; ORG, ocean ridge granite.

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**Fig. 16.** Comparison of detrital zircon ages from the Asís Lithodeme with the Cosoltepec Formation (Keppie et al. 2006), and potential source regions (modified after Murphy et al. 1999).
Neoproterozoic and subsequently recycled during Palaeozoic tectonic events (Murphy et al. 2000, 2004). This situation may be analogous to Mesozoic–Cenozoic terrane accretion in the Canadian Cordillera, where recent studies have shown that such accretion can be thick skinned, incorporating the mantle lithosphere to a depth of c. 150 km (McKenzie et al. 2005).

Granitoid rocks

Field relationships (Middleton et al. 2005) and similar geochemical and isotopic characteristics suggest that the mylonitic granite is the deformed equivalent of the c. 470–420 Ma megacrustic granitoids. Their 1.5–1.9 Ga, DTSa ages are very similar to those of the c. 1 Ga basement of Oaxaquia, which are: (1) 1.4–1.6 Ga for metaigneous rocks; (2) 1.5–2.0 Ga for the metasedimentary rocks (Patchett & Ruiz 1987; Ruiz et al. 1988; Weber & Köhler 1999). These characteristics suggest that the granitoid magma was derived by crustal anatexis from a basement similar to Oaxaquia, which is inferred to underlie the Acatlán Complex (Keppie 2004). Their geochemistry is very similar to that of Ordovician undeformed granitoid plutons in the northern Acatlán Complex (Miller et al. 2005). Although these geochronological data do not sufficiently resolve the relative timing of basaltic and granitoid magmatism in the Asís Lithodeme, regional considerations are consistent with the interpretation that the amphibolites and the granitoids in the Asís Lithodeme may represent a bimodal suite.

Summary and conclusions

According to Talavera-Mendoza et al. (2005), the Acatlán Complex reflects c. 480–470 Ma SE-vergent subduction, and high-pressure (eclogite- and blueschist-facies metamorphism) during which time the Piaxtla Suite accreted to Laurentia. If correct, the Acatlán Complex would have originated within the Iapetus Ocean. However, a concordant U–Pb zircon (isotope dilution thermal ionization mass spectrometry) age of 346 ± 3 Ma from a Piaxtla Suite eclogite and c. 345 Ma SHRIMP analysis from a Piaxtla Suite migmatite (Middleton et al. 2005) indicate that the Asís Lithodeme underwent eclogite-facies metamorphism followed by rapid exhumation in the Carboniferous (Middleton et al. 2005), implying that the Asís protoliths could not have formed within the Iapetus Ocean, which closed in the Silurian.

The lithologies of the Asís area could have formed along the Gondwanan flank of the Rheic Ocean during the rifting of Avalonia from Gondwana. The combination of within-plate continental tholeiitic mafic rocks, crustally derived granitoids, and continentally derived clastic rocks suggests that the Asís Lithodeme protoliths represent a rift–passive margin sequence. Although the age constraints on the time of deposition of the metasedimentary rocks are rather wide (c. 700–420 Ma), the Tremadoc age of the adjacent Tinú Formation provides a minimum age for initiation of the passive margin sequence. As Oaxaquia remained along the Gondwanan margin until Carboniferous collision with Laurussia, the data suggest that some terrane rifted from the Oaxacan portion of the Gondwanan margin in the Tremadoc. Most recent palinspastic reconstructions place Avalonia adjacent to Oaxaquia and Amazonia in the late Neoproterozoic (e.g. Keppie et al. 2003b, and references therein). Although Prigmore et al. (1997) inferred that Avalonia separated from Gondwana in the Late Cambrian to early Tremadoc based upon subsidence curves, they documented three episodes of rapid subsidence in Avalonia: Early Cambrian, Late Cambrian–Tremadoc, and Late Ordovician. Landing (1996, 2004) has suggested that the distinct Avalonian fauna indicates rifting and separation of Avalonia from Gondwana in the latest Neoproterozoic to Early Cambrian, which is consistent with the Early Cambrian rapid subsidence recorded in Avalonia (Prigmore et al. 1997). However, Avalonian fauna gradually become indistinguishable from Gondwanan fauna in the Late Cambrian and Early Ordovician (Fortey & Cocks 2003; Landing 2005), suggesting that Early Cambrian rifting resulted in the development of a narrow seaway and that Avalonia essentially retained its peri-Gondwanan location. Keppie et al. (2003b) proposed that late Neoproterozoic collision of an ocean ridge with the subducting northern margin of Gondwana produced a Baja California type of margin, resulting in transtensional rifting and separation of Avalonia in the Cambrian. It is inferred that Avalonia moved along the Gondwanan margin with the fauna gradually managing to cross the small ocean barrier.

Faunal and palaeomagnetic data (e.g. Cocks & Torsvik 2002; Fortey & Cocks 2003) indicate that Avalonia drifted about 1800 km north of the Gondwanan margin between 485 and 460 Ma (Hamilton & Murphy 2004), ending its peri-Gondwanan affinity (Fig. 17). This drift of Avalonia from the Gondwanan margin resulted in the formation of the Rheic Ocean, and corresponds to the second phase of subsidence documented by Prigmore et al. (1997).

The data presented herein, however, cannot distinguish between the various mechanisms proposed for the origin of the Rheic Ocean. According to Van Staal et al. (1998), the Rheic
Ocean started as a back-arc basin, but evidence for arc-related rocks coeval with rifting along the Gondwanan margin is equivocal. In this case, the protoliths of the Asís Lithodeme may have formed on a continental margin inboard of a rifted arc.

As the opening of the Rheic Ocean is coeval with the onset of NW-directed subduction and ridge–trench collision along the Laurentian margin (see Van Staal et al. 1998, Stampfli & Borel 2002), the origin of the Rheic Ocean may be geodynamically linked to slab pull (Murphy et al. 2006) in a manner analogous to the opening of Neotethys in the Cenozoic (Stampfli & Borel 2002). In this scenario the protoliths of the Asís Lithodeme may have formed in a continental rift environment.

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