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**Química mineral, isótopos de Sr-Nd-Os e
elementos do grupo da platina dos
Peridotitos do vulcão Los Gemelos, Meseta
de Canquel, Patagônia Argentina**

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RESUMO

Os xenólitos mantélicos da Meseta de Canquel, Patagônia, são objeto deste estudo e compreendem a primeira investigação dos espinélio-peridotitos datados do Mesoproterozoico (1,3–0,9 Ga). A composição dos xenólitos varia de harzburgitos a lherzolitos, com baixo teor de clinopiroxênio, indicando eventos de fusão parcial. Os baixos teores de Al_2O_3 e CaO em rocha total (ambos <1.5 wt.%), o alto Mg# nas fases silicáticas (90–95) e o espinélio rico em Cr (Cr# 0.20–0.42), juntamente com a depleção de elementos altamente siderófilo (Re, Ru, Pd) em relação ao manto superior primitivo, o que confirma as altas taxas de fusão e o caráter fortemente depletado das amostras analisadas. A presença de flogopita, associada a presença abundante de veios de composição alcalina com alta e baixa SiO_2 , são evidências incontestáveis de metassomatismo modal. O enriquecimento do clinopiroxênios em elementos terras raras leves (LREE), assim como o padrão em U dos REE da rocha total, são evidências de metassomatismo críptico. A interação entre a rocha e o líquido (*melt*) é evidenciada pelo conteúdo de elementos maiores (e.g., Mg# vs. elementos basálticos) e traços (e.g., La/Yb_N vs. Sr/Y e Ti/Eu; e Sr vs. Ti/Eu) dos piroxênios. Além disso, as composições dos líquidos em equilíbrio com os clinopiroxênios coexistentes com a flogopita, assim com as razões isotópicas de rocha total de Sr-Nd, sugerem a interação com componentes relacionados a subducção. Os efeitos do metassomatismo são impressos mais proeminentemente nas porções mantélicas profundas (~70 km) em comparação com as porções mais rasas (~50 km). Tais características são consistentes com a hibridização do manto litosférico depletado com líquidos alcalinos ricos em K. Considerando o sistema de subducção de longa duração estabelecido na região de *back-arc* da Patagônia durante o Fanerozoico, é muito provável que o evento metassomático registrado pelos xenólitos mantélicos de Los Gemelos esteja associado ao soerguimento de material através de uma janela astenosférica durante o Paleoceno-Eoceno, que causou a fusão parcial da placa oceânica.

Palavras-chave: Espinélio Peridotitos; Fusão Parcial; Mesoproterozoico, Metassomatismo mantélico por fluidos ricos em K; Isótopos de Sr, Nd e Os; Manto da Patagônia.

ABSTRACT

Mantle xenoliths from Canquel Plateau, Patagônia, are object of this study, which represent the first investigation of a new spinel-bearing suite dated from Mesoproterozoic (1.3–0.9 Ga). The composition vary between harzburgites and clinopyroxene-poor lherzolites characterized by typical indicators of partial melt extraction, such as low whole-rock Al_2O_3 and CaO contents (both <1.5 wt.%), high Mg# of silicate phases (90–95), and Cr-rich spinel (Cr# 0.20–0.42). The depletion of incompatible highly siderophile elements (Re, Ru, Pd) relative to the primitive upper mantle certify the high-degrees of melt extraction and the highly-depleted character of the samples. The existence of phlogopite, accompanied by abundant alkaline glass veins containing high- and low- SiO_2 compositions are undoubtable evidence of modal metasomatism. The light rare earth element (LREE) enrichment observed in clinopyroxene, as well as the whole-rock U-shaped REE pattern are evidence of cryptic metasomatism. The melt-rock interaction is corroborated by major (i.e., Mg# vs. basaltic elements) and trace (i.e., La/Yb_N vs. Sr/Y and Ti/Eu ; and Sr vs. Ti/Eu) element contents of pyroxenes. Moreover, the calculated compositions of melts in equilibrium with clinopyroxene coexisting with phlogopite, as well as whole-rock Sr–Nd isotopic ratios suggest interaction with subduction-related components. Remarkably, the metasomatic overprint is selectively prominent in the deeper (~70 km) mantle column compared to the shallower (~50 km). These features are consistent with the hybridization of the depleted lithosphere with a K-rich alkaline melt. Considering the long-lived subduction system established in the Patagonia back-arc throughout the Phanerozoic, it is most likely that the metasomatic event recorded by Los Gemelos mantle xenoliths is associated with the upwelling of asthenospheric material through a slab-window during the Paleocene-Eocene, which promoted the partial melting of the oceanic crust.

Keywords: Spinel peridotites; Partial melting; Mesoproterozoic; Hydrous and high-K mantle metasomatism; Sr, Nd and Os isotopes; Patagonian mantle

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CAPÍTULO 1

INTRODUÇÃO

1. INTRODUÇÃO

1.1. Objetivos

Esse trabalho discute aspectos relacionados à evolução do manto litosférico subcontinental (SCLM) abaixo do vulcão Los Gemelos, localizado na Meseta de Canquel, Patagônia Argentina. Ele almeja contribuir com o melhor entendimento dos processos e agentes metassomáticos envolvidos na heterogeneidade mineralógica e química dos xenólitos mantélicos trazidos à superfície no Pleistoceno a partir do magmatismo alcalino. Para tanto, realizamos a caracterização petrográfica e produzimos resultados de: (1) química mineral em olivina, ortopiroxênio, clinopiroxênio, espinélio, flogopita e veios metassomáticos, (2) geoquímica de rocha total, (3) isótopos de Sr-Nd-Os, e (4) elementos altamente siderófilos (elementos do grupo da platina e Re) para 16 amostras de peridotitos anidros e hidratados.

1.2. Caracterização do problema

O estudo em detalhe dos xenólitos mantélicos de Los Gemelos se faz necessário pelo seu ineditismo e devido à excentricidade mineralógica e composicional observada nessas amostras. Elas registram a percolação de líquidos de composição adakítica, que produziu veios em diversas amostras, além de flogopita em duas delas. Ambos os exemplos definem metassomatismo modal, que inevitavelmente promove metassomatismo críptico. Na região de *back-arc* da Patagônia, metassomatismo adakítico foi reportado apenas para Cerro del Faile (Killian & Stern, 2002; Faccini *et al.*, 2013) e para a região de Paso de Indios (Bertotto *et al.*, 2021). Já a existência de flogopita está restrita aos xenólitos mantélicos de Gobernador Gregores (Gorring & Kay, 2001; Laurora *et al.*, 2001; Rivalenti *et al.*, 2004) e do Campo Vulcânico de Pali-Aike (Stern *et al.*, 1999; Gervasoni *et al.*, 2012; Zaffarana *et al.*, 2014; Jalowitzki *et al.*, 2016).

1.3. Contexto geológico

A Cordilheira dos Andes é dividida em quatro zonas vulcânicas, com seus limites definidos por meio da taxa de convergência e ângulo de mergulho das placas subductantes. De acordo com Stern (2004), elas são: Zona Vulcânica Norte (ZVN; 5°N-2°S), Zona Vulcânica

Central (ZVC; 14°-27°S), Zona Vulcânica Sul (ZVS; 33°-46°S) e Zona Vulcânica Austral (ZVA; 49°-55°S). As ZVS e ZVA são separadas por um gap vulcânico (46-49°S; [Stern, 2004](#)) que corresponde à colisão da Dorsal do Chile com a margem oeste da placa Sul-americana, formando uma junção tríplice onde a placa de Nazca está à norte e a da Antártica à sul. A colisão desta dorsal resultou na abertura de uma janela astenosférica que favoreceu a fusão parcial do manto litosférico por descompressão, gerando basaltos intraplaca com características geoquímicas e isotópicas de basaltos de ilhas oceânicas (OIBs) na região de *back-arc* ([Ramos & Kay, 1992](#); [Gorring et al., 1997](#)).

O magmatismo de *back-arc* na região centro-norte da Patagônia, que inclui as mesetas (ou plateaus) de Somún Curá e a de Canquel, é composto essencialmente por lavas geradas do Oligoceno ao Pleistoceno, com pico efusivo em 23 ± 5 Ma ([Jalowitzki et al., in prep.](#)). Ele é subdividido em sequência plateau (Oligoceno-Mioceno) e post-plateau (Mioceno superior-Pleistoceno) ([Ramos & Kay, 1992](#); [Gorring et al., 1997](#); [Jalowitzki et al., in prep.](#)). A Meseta de Canquel (Figura 1) representa um dos principais eventos magmáticos de platô da Patagônia, sendo caracterizada pela ocorrência de derrames basálticos de idade Cenozoica ([Anselmi et al., 2004](#); [Alric & Massaferro, 2017](#)). Estes derrames são essencialmente compostos por olivina basaltos da Formação Sierra Cuadrada ([Haller et al., 2020](#)), que registra o início da atividade vulcânica no Eoceno (a partir de 28 Ma), mas apresenta idades predominantes do Mioceno (23-21 Ma) ([Anselmi et al., 2004](#) e referências nele contidas). No extremo norte da meseta, a parte basal dos derrames é composta de lavas básicas alcalinas e por corpos de gabros e diabásios da Formação El Canquel, que datam do Eoceno (51-45 Ma; [Anselmi et al., 2004](#) e referências nele contidas). Cones vulcânicos pouco dissecados, representando uma atividade vulcânica do tipo pós-platô, estão localizados acima da meseta e pertencem a Formação Sierra Cuadrada ([Anselmi et al., 2004](#) e referências nele contidas).

A estratigrafia próxima ao vulcão Los Gemelos, na porção nordeste da Meseta de Canquel (Figura 1), é constituída por traquiandesitos basálticos (base) a basanitos (topo) ([Alric & Massaferro, 2017](#)). Segundo estes autores, o aumento de alcalinidade das rochas em direção ao topo indica a diminuição da taxa de fusão parcial da fonte, característica dos basaltos do estágio pós-platô. Embora haja o registro de abundantes ocorrências de xenólitos mantélicos na Patagônia (36°S - 52°S), a descoberta desta nova suíte é inédita na região da Meseta de Canquel. Os xenólitos em questão são classificados como espinélio-lherzolitos.

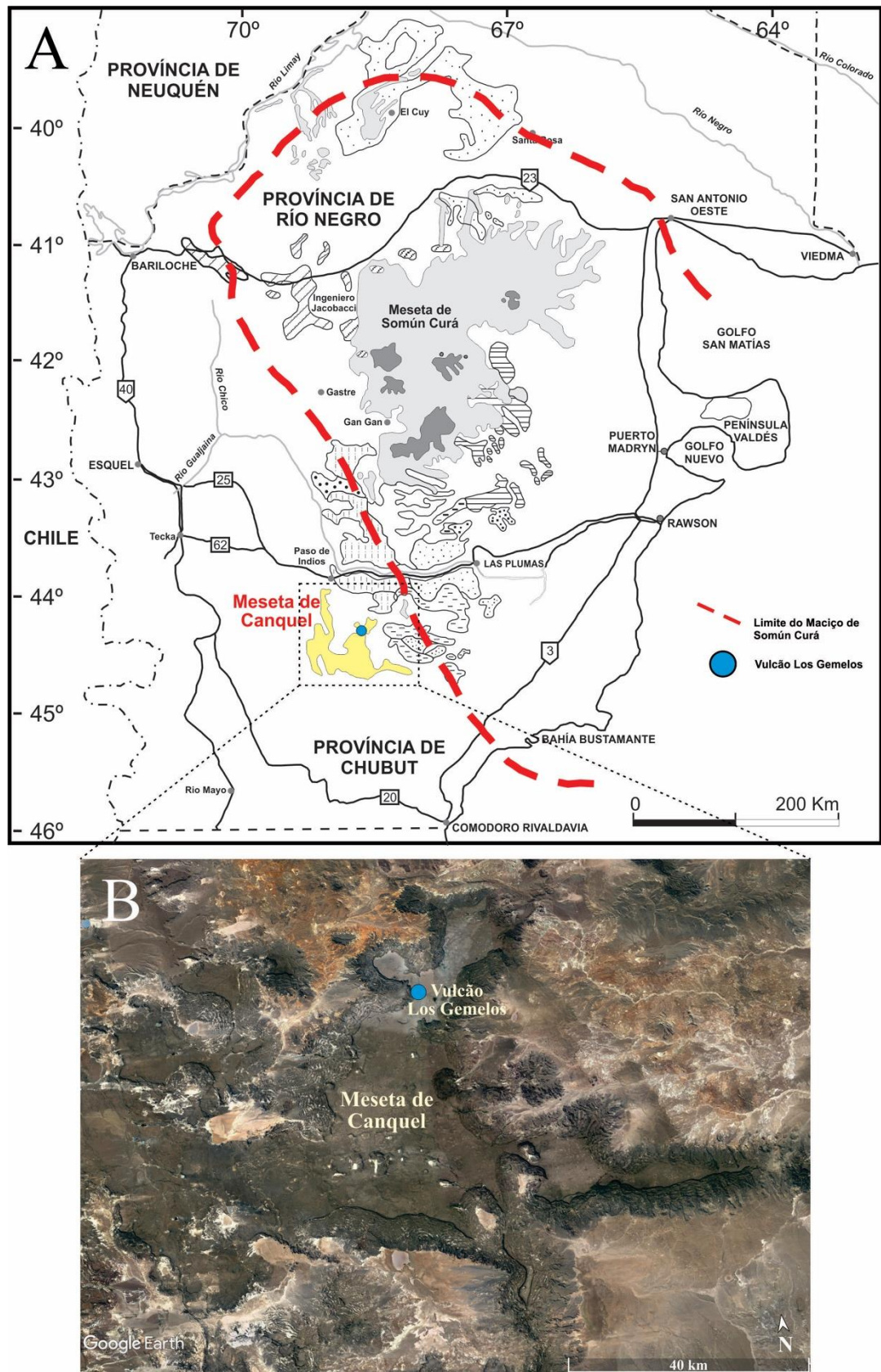


Figura 1. (A) Mapa de localização da área de estudo mostrando o vulcão Los Gemelos no contexto dos maiores platôs basálticos da região centro-norte da Patagônia (Somún Curá e Canquel). (B) Mapa da Meseta de Canquel mostrando o vulcão Los Gemelos (Fonte Google Earth).

1.4. Xenólitos mantélicos no contexto patagônico

Os xenólitos são fragmentos diretos do manto transportados até a superfície de maneira rápida e abrupta por eventos vulcânicos, especialmente aqueles de natureza alcalina. A paragénese mineralógica típica de xenólitos mantélicos provenientes da litosfera em ambientes de *back-arc* continental é constituída por olivina (ol), ortopiroxênio (opx), clinopiroxênio (cpx) e espinélio (sp). Esta assembleia mineral é conhecida como sp-peridotito (~40 a ~80 km). Ainda, peridotitos oriundos da zona de estabilidade da granada são estáveis em profundidades superiores a 80 km, estando o limite litosfera-astenosfera usualmente em 100 km. No entanto, peridotitos contendo plagioclásio como fase aluminosa são raros (estável até ~35 km). Embora essas amostras forneçam valiosas informações acerca da heterogeneidade mineralógica e química do manto litosférico, sua natureza e a influência de componentes relacionados às zonas de subducção ou de líquidos de origem astenosférica permanecem em debate.

Os estudos mais relevantes envolvendo amostras de xenólitos mantélicos da Patagônia foram produzidos para os vulcões: Agua Poca (Bertotto, 2000; Jalowitzki *et al.*, 2010; Bertotto *et al.*, 2013); Huanul (Bertotto *et al.*, 2021); Cerro el Mojón e Estância Alvarez (Mallmann, 2004); Prahuanियeu (Bjerg *et al.*, 2009); Cerro Chenque, Cerro Matilde e Cerro León (região de Paso de Indios: Bertotto *et al.*, 2022); Cerro de los Chenques (Rivalenti *et al.*, 2007; Dantas *et al.*, 2009); Tres Lagos (Ntaflos *et al.*, 2007; Mazzucchelli *et al.*, 2016); Cerro del Fraile (Kilian & Stern, 2002; Faccini *et al.*, 2013); Gobernador Gregores (Gorring & Kay, 2001; Laurora *et al.*, 2001; Rivalenti *et al.*, 2004; Zaffarana *et al.*, 2014; Jalowitzki *et al.*, 2016); Pali-Aike (Stern *et al.*, 1999; Gervasoni *et al.*, 2012; Zaffarana *et al.*, 2014; Jalowitzki *et al.*, 2016); e Coyhaique (Jalowitzki *et al.*, 2017; Novais-Rodrigues *et al.*, 2021). Por outro lado, estudos regionais discutiram a petrografia, geoquímica e isótopos de diferentes localidades (ex. Rivalenti *et al.*, 2004; Bjerg *et al.*, 2005; Conceição *et al.*, 2005; Schilling *et al.*, 2008, 2017; Mundl *et al.*, 2015).

Embora diversos estudos tenham sido realizados, a ocorrência de xenólitos mantélicos no *back-arc* patagônico com feições petrográficas similares às observadas em Los Gemelos são raras. Por exemplo, a cristalização de flogopita, mineral hidratado rico em K, foi descrita no Campo Vulcânico de Pali-Aike e em Gobernador Gregores (Stern *et al.*, 1999; Gorring & Kay, 2000; Laurora *et al.*, 2001; Rivalenti *et al.*, 2004; Gervasoni *et al.*, 2012; Zaffarana *et al.*,

2014; Jalowitzki *et al.*, 2016). Da mesma forma, metassomatismo adakítico é descrito em xenólitos mantélicos de Cerro del Fraile (Kilian & Stern, 2002; Faccini *et al.*, 2013) e em Paso de Indios (Bertotto *et al.*, 2021). Com relação aos processos metassomáticos envolvidos no enriquecimento do SCLM da Patagônia, os mais relevantes estão relacionados à percolação de fluidos/melts derivados da placa oceânica subductante (Coyhaique, Cerro el Mojón, Estância Alvarez). A percolação de líquidos de origem astenosférica tem sido invocada para explicar o metassomatismo registrado por xenólitos de Prahuanियeu, Cerro de los Chenques e Pali-Aike. Metassomatismo carbonatítico foi atribuído ao SCLM abaixo de Gobernador Gregores. Por outro lado, o manto depletado como resposta à eventos de fusão parcial está exemplificado por Agua Poca, Huanul e Tres Lagos.

1.5. Depleção e posterior enriquecimento do manto terrestre

Ao longo do tempo geológico, o manto terrestre tem registrado uma complexa história de depleção e enriquecimento em resposta a eventos de fusão parcial e metassomatismo, respectivamente. As heterogeneidades observadas em resíduos sólidos (xenólitos mantélicos), por vezes, não são facilmente explicadas pela fusão parcial, exigindo a participação de um manto enriquecido devido a eventos metassomáticos que desempenham papel importante na composição mineralógica (litológica) e química dessas rochas. Eventos metassomáticos desempenham papel importante na composição mineralogia e na química do manto, estando o metassomatismo modal relacionado à ocorrência de fases hidratadas (ex. pargasita e flogopita), enquanto o metassomatismo críptico está relacionado à interação com fluidos ou líquidos (*melts*) enriquecidos em elementos incompatíveis. Nesse contexto, a fusão parcial é tipicamente marcada pelo Mg# elevado nas fases silicáticas do manto (>90 , onde $Mg\# = Mg/(Mg+Fe)$), baixa concentração de elementos basálticos (ex. Al_2O_3 e CaO) na rocha total e depleção de elementos terras raras leves (LREE) com relação aos pesados (HREE) na rocha total e em suas fases minerais. Por outro lado, o metassomatismo é marcado pelo enriquecimento em elementos altamente incompatíveis com relação aos compatíveis (ex. LREE/HREE).

1.5.1. Metassomatismo por fluidos aquosos e líquidos silicáticos

A interação entre fluidos aquosos e/ou líquidos silicáticos com xenólitos mantélicos promove heterogeneidade, que depende da composição do agente metassomático, da natureza do manto (normalmente empobrecido), além da profundidade ([Rampone *et al.*, 2008](#)). Líquidos silicáticos têm ampla variação química, como exemplificado pelos magmas hospedeiros de xenólitos mantélicos (ex. basaltos alcalinos, kimberlitos, kamafugitos, nefelinitos e melilitos). Comumente, a distinção entre líquido silicático e fluido aquoso não é simples, pois a existência de fases hidratadas como flogopita e anfibólio não é conclusiva acerca da interação do manto com fluido aquoso, já que o líquido silicático pode conter composição hidratada. Como exemplo, a fusão da placa oceânica eclogitizada e hidratada tende a gerar líquidos de afinidade cálcio-alcalina, tais como dacito e andesito, ou adakito. Nesse sentido, cabe observar que fluidos aquosos usualmente não têm afinidade com elementos de alta carga (HFSE: Nb, Ta, Zr, Hf, Ti), mas são enriquecidos em elementos litófilos de raio iônico grande (LILE) altamente incompatíveis e móveis (ex. Rb, Ba, K, Pb, Sr). Portanto, a comparação entre as abundâncias de LILE e HFSE pode auxiliar na discriminação entre os dois agentes metassomáticos. Assim, elevadas razões LILE/HFSE sugerem a interação com um fluido aquoso, enquanto elevadas concentrações de HFSE sugerem a influência de um líquido silicático.

1.5.2. Adakitos

De acordo com [Defant & Drummond \(1990\)](#), o primeiro registro de rochas adakíticas foi documentado na Ilha de Adak, no Alasca, nas ilhas Aleutas, para designar andesitos e dacito. O termo adakito é utilizado para descrever rochas vulcânicas ou intrusivas datadas do Cenozoico, as quais estão associadas ao processo de subducção de litosfera oceânica jovem (≤ 25 Ma). Nesse processo, a placa oceânica e eventualmente os sedimentos sobrejacentes fundem devido a uma fonte de calor. Ao contrário da maioria dos magmas cálcio-alcalinos de arco (basaltos, andesitos, dacitos e riolitos), formados em decorrência da fusão do manto peridotítico metassomatizado por fluidos ([Gill, 1981](#)).

Adakitos são caracterizados por altos conteúdos de SiO_2 ($>56\%$) e Al_2O_3 ($>15\%$), com baixo MgO ($<3\%$, raramente $>6\%$). Essas rochas também são caracterizadas por baixas

concentrações de HFSE e por razões $^{87}\text{Sr}/^{86}\text{Sr}$ geralmente inferiores a 0,7040 (Defant & Drummond, 1990; Martin *et al.*, 2005; Castillo, 2006). Ainda, devido às baixas concentrações de Y e Yb (<15 e 1,9 ppm, respectivamente), acompanhadas de alto Sr (>300 ppm), as razões Sr/Y e La/Yb são altas (>20). Por outro lado, rochas cálcio-alcálinas apresentam conteúdo de SiO_2 próximo ao observado em adakitos (~52 a 71%), já a composição em elementos de alto potencial iônico tais como Zr, Y e Nb são mais elevados (Ruggiero & Oliveira, 2010).

Sendo assim, as características geoquímicas dos adakitos são consistentes com a fusão parcial da crosta oceânica de composição basáltica metamorfizada nas fácies anfibolito (com granada) ou eclogito. Assim, a depleção de elementos terras raras pesados (ex. Yb) e de Y é justificada pela seletiva retenção desses elementos na placa oceânica devido aos seus elevados coeficientes de partição para granada. Ainda, o sistema de subducção favorece o enriquecimento de Sr e La, justificando, portanto, as elevadas razões Sr/Y e La/Yb dos adakitos, em concordância com o trabalho de Ruggiero & Oliveira (2010) ao qual diferenciou suítes vulcânicas adakíticas e cálcio-alcálinas com base na razão de Sr/Y.

1.6. Heterogeneidade mantélica marcada por isótopos

1.6.1. Isótopos de Re-Os

De acordo com Huston *et al.* (2023), Re e Os são elementos móveis em condições oxidantes, como ambientes associados a arcos vulcânicos. O particionamento dos elementos altamente siderófilos (HSE; elementos do grupo da platina - PGEs + Re) durante o processo de fusão é controlado principalmente pela presença ou ausência de sulfetos na fonte e, em menor proporção, pelas fases silicáticas e óxidos (Brenan *et al.*, 2016). Durante o processo de fusão parcial do manto, o Re se comporta como elemento moderadamente incompatível, sendo concentrado na fase líquida e, portanto, progressivamente enriquecido na crosta terrestre. Já o Os se comporta como elemento compatível, sendo retido no resíduo depleto da fusão. Sendo assim, a crosta continental possui composição isotópica de $^{187}\text{Os}/^{188}\text{Os}$ altamente radiogênica, refletindo sua elevada razão $^{187}\text{Re}/^{188}\text{Os}$ (Fonseca *et al.*, 2007; Brenan, 2008). No entanto, o comportamento do Re, Os e Pt no manto difere de outros sistemas isotópicos comumente utilizados, baseados em elementos litófilos incompatíveis, devido à ausência de fracionamento pronunciado por fusões mantélicas de pequena a moderada magnitude (Huston *et al.*, 2023). Sendo assim, a composição isotópica de Os do manto fértil, moderadamente empobrecido, é

coerente com a evolução de longo prazo em proporções quase condriticas de Re/Os (Meisel *et al.*, 2001; Gannoun *et al.*, 2007).

O sistema Re-Os é baseado no decaimento radioativo do ^{187}Re para ^{187}Os , o qual apresenta meia vida de aproximadamente 42,3 Ga, de acordo com Smoliar *et al.* (1996), e tem por objetivo calcular idades de eventos que causaram perturbações ao equilíbrio do sistema, pós formação. Dessa forma, dois modelos de idade podem ser calculados para o sistema isotópico Re-Os, sendo a idade modelo do manto (T_{MA}) e a idade modelo de empobrecimento de Re (T_{RD}), ambas pressupõem que a amostra foi diretamente derivada do manto e, portanto, são aplicadas mais comumente a materiais como xenólitos mantélicos e basaltos.

Idades modelo T_{MA} são análogas aos modelos utilizados em outros sistemas de decaimento (e.g. T_{DM} ; Sm e Nd), refletindo a idade de separação do magma, ou o resíduo dos processos de convecção mantélica (composicionalmente semelhante ao manto primitivo superior - PUM), assumindo que o fragmento amostrado não passou por nenhum outro evento de homogeneização ou convecção, permanecendo isolado desde então. Já a idade T_{RD} reflete o último evento de depleção ao qual a rocha foi submetida, para tanto, é preciso assumir que o evento de fusão parcial foi forte o suficiente (>20% de fusão; Mundl *et al.*, 2016; Luguet & Pearson, 2019) para extrair todo o Re, deixando o peridotito livre de qualquer resíduo de Re (Walker *et al.*, 1989). Sendo assim, a razão $^{187}\text{Os}/^{188}\text{Os}$ calculada é idêntica a razão no momento do evento de depleção e as idades T_{MA} e T_{RD} serão idênticas. As idades T_{RD} serão sistematicamente mais jovens que as idades T_{MA} , quanto menor for a taxa de fusão parcial ao qual a rocha fora submetida, como um lherzolito (Luguet & Pearson, 2019).

1.6.1.1. Elementos altamente siderófilos (PGE + Re)

Os elementos do grupo da platina, incluindo Rh, Ru, Pd, Os, Ir e Pt, apresentam resistência às reações químicas e permanecem estáveis em sua forma metálica. Além disso, eles podem ser subdivididos em dois grupos distintos: Ir-PGE (Os, Ir, Ru) e Pd-PGE (Pd, Pt, Rh). Esses elementos também exibem uma forte afinidade com o Fe, tornando-os altamente siderófilos (HSE), juntamente com o Re.

Em rochas do manto, os HSE são predominantemente encontrados em sulfetos de metais básicos (Alard *et al.*, 2000; Luguet *et al.*, 2001). Contudo, em níveis elevados de fusão parcial (>20 %), esses sulfetos devem ser completamente consumidos no líquido (Luguet *et al.*, 2003; Mundl *et al.*, 2016). Isso resulta em uma forte diminuição da concentração de

elementos do grupo Pd-PGE e do Re no manto residual (e em menor escala de Pt), uma vez que esses elementos são mais incompatíveis em relação aos elementos do grupo Ir-PGE (Pearson *et al.*, 2004). Os HSE remanescentes, ou seja, Os, Ir e Ru (além da Pt), são subsequentemente redistribuídos. As fases hospedeiras específicas no manto terrestre onde os PGE's residem ainda não são totalmente compreendidas. No entanto, de acordo com Mitchell & Keays (1981), eles parecem estar concentrados em minerais na seguinte ordem: granada < olivina < ortopiroxênio < clinopiroxênio < espinélio.

1.6.2. Isótopos de Rb-Sr e Sm-Nd: reservatórios mantélicos

Em função da ausência ou escassez de materiais mantélicos diretos, a análise de basaltos oceânicos tornou-se uma fonte de dados acessível. Devido a isso, o comportamento isotópico dos basaltos oceânicos revela tendências que corroboram com o estabelecimento de membros finais ou reservatórios mantélicos, os quais são denominados DM (manto empobrecido), HIMU (high- μ , onde μ = elevada razão $^{238}\text{U}/^{204}\text{Pb}$), EMI e EMII (mantos enriquecidos), e PREMA (manto prevalente) (Zindler & Hart, 1986; Hofmann, 1997). Aqui não abordaremos o reservatório HIMU, pois ele é definido a partir dos isótopos de chumbo (Pb) e não dispomos desses resultados.

1.6.2.1. Manto empobrecido (do tipo MORB)

O mecanismo principal que impulsiona a formação desse reservatório é a fusão parcial do manto, que promove a formação da crosta. Como resposta a esse processo, os elementos incompatíveis concentram no líquido (magma), enquanto os compatíveis tendem a permanecer no resíduo sólido, de elementos mais compatíveis. Devido às baixas razões $^{87}\text{Sr}/^{86}\text{Sr}$ e elevadas razões $^{143}\text{Nd}/^{144}\text{Nd}$, esse reservatório é definido como um membro final extremo com caráter isotópico empobrecido (Hart, 1988).

1.6.2.2. Manto enriquecido (EMI e EMII)

Os reservatórios mantélicos enriquecidos, se referem a níveis elevados de elementos incompatíveis e características isotópicas distintas, o que permite a separação em pelo menos dois membros finais. O EMI possui razões moderadas e o EMII elevadas razões de $^{87}\text{Sr}/^{86}\text{Sr}$, ambos com baixas razões de $^{143}\text{Nd}/^{144}\text{Nd}$. Os reservatórios mais prováveis para a composição

do EMI incluem a litosfera oceânica antiga, muitas vezes acompanhada por sedimentos pelágicos, que passou por processos de reciclagem ao longo do tempo ([Hart *et al.*, 1992](#); [Hauri *et al.*, 1994](#)). Ademais, o manto inferior metassomatizado também pode ser uma fonte em potencial ([Collerson & Kamber, 1999](#)). Já o manto enriquecido II apresenta razões isotópicas que se aproximam das médias observadas na crosta continental superior ou em sedimentos continentais terrígenos que estão atualmente submetidos a processos de subducção ([Stracke, 2012](#); [Stracke *et al.*, 2005](#)).

CAPÍTULO 2

METODOLOGIA

2. METODOLOGIA

2.1.Preparação das amostras

Dezesseis lâminas petrográficas polidas foram confeccionadas para os xenólitos de Los Gemelos. Devido ao tamanho reduzido dos xenólitos, 9 amostras foram selecionadas e fragmentadas através do SELFRAG® no Laboratório de Geocronologia e Geoquímica Isotópica, Instituto de Geociências (IG), da Universidade de Brasília (UnB). Embora as amostras sejam frescas, minerais alterados foram removidos das amostras desagregadas com auxílio de uma lupa estereoscópica binocular de bancada. Após, a pulverização foi executada através do uso de um cadinho e pistilo de ágata com o objetivo de gerar pó de rocha (rocha total) com granulometria inferior a 200 mesh (fração talco) para a viabilização das análises geoquímicas e isotópicas.

2.2.Petrografia

O propósito da descrição petrográfica consiste na identificação da paragênese mineral e das principais texturas presentes nos peridotitos, com a assistência de um microscópio óptico petrográfico, do Laboratório de Microscopia, localizado nas instalações do IG-UnB. A quantificação das fases foi realizada digitalmente, com o auxílio do software JMicroVision ([Roduit, 2007](#)), com a contagem de, no mínimo, 2000 pontos por lâmina.

2.3.Microsonda eletrônica

Os elementos maiores das fases minerais (olivina, ortopiroxênio, clinopiroxênio, espinélio e flogopita) e dos veios das 16 amostras de xenólitos de Los Gemelos foram determinados por meio de dispersão de comprimento de onda (WDS) na microsonda eletrônica JEOL JXA-8230 (EPMA) instalada no Laboratório de Microsonda Eletrônica do IG-UnB. As análises foram realizadas com uma voltagem de aceleração de 15 kV, um feixe de 10 nA e um diâmetro de feixe de 1 µm no modo de ponto. Para as medições de flogopita, utilizamos um feixe com 5 µm. O tempo de contagem no pico foi de 10 s por pico analítico e 5 s para o *background*. Todos os resultados foram corrigidos utilizando um programa de correção ZAF: Z (número atômico), A (probabilidade de absorção), F (fluorescência de raios-

X secundária), desenvolvido internamente. A calibração do equipamento foi feita a partir dos seguintes padrões: andradita (CaO e FeO), albita (Na₂O), forsterita (MgO), microclínio (K₂O, Al₂O₃ e SiO₂), pirofanita (TiO₂ e MnO), óxido de cromo (Cr₂O₃), óxido de níquel (NiO), topázio (F) e vanadinita (V₂O₃ e Cl).

2.4.Elementos traço por LA-ICP-MS

Os elementos traço e terras raras foram determinados para 10 amostras no Laboratório do Centro Interdipartimentale Grandi Strumenti (CIGS), da Università di Modena e Reggio Emilia, Itália. Os resultados foram gerados em um espectrômetro de massa Thermo Fisher Scientific XSERIES-2 acoplado a um sistema de ablação a laser New Wave UP 213. A redução dos dados foi realizada com o software PlasmaLab® da Thermo Fisher Scientific, utilizando os padrões externos NIST610 (SD <4,0 %) ([Jochum et al., 2005](#)) e BCR-2 (SD <3,8 %) ([Jochum et al., 2015](#)).

As análises foram realizadas com o feixe do laser de 55 µm, potência de saída de 70 % em peso, taxa de repetição de 10 Hz, tempo de permanência de 30 s, e fluência do feixe de laser de 8 J/cm². A superfície da amostra foi inicialmente submetida a uma pré-ablação com feixe do laser de 65 µm, potência de 45 % em peso e taxa de repetição de 10 Hz. O isótopo ⁴⁴Ca foi usado como padrão interno para o clinopiroxênio, enquanto o isótopo ²⁹Si foi utilizado para a flogopita.

2.5.Geoquímica de rocha total e isótopos radiogênicos de Sr-Nd

A preparação das amostras, a geoquímica de rocha total e as análises isotópicas de Sr-Nd foram realizadas nos laboratórios de geoquímica isotópica da Universidade de Tübingen, na Alemanha. A perda ao fogo (LOI) foi determinada colocando uma amostra de 100 mg em um forno a 1050 °C por uma hora, resfriando em uma dessecadora e pesando novamente. Devido à quantidade limitada de material da amostra, um espectrômetro de massa com plasma acoplado (ICP-MS) ThermoFisher Scientific iCAP-Qc foi utilizado para determinar as composições de elementos maiores e traços.

2.5.1. *Elementos Maiores*

Para análise de elementos maiores, uma amostra de 100 mg do material foi misturada em um cadinho de platina com 500 mg de Merck© Spectromelt® A 12 (mistura de 66 % de tetraborato de dilítio e 34 % de metaborato de lítio) e fundida a 1200 °C, formando uma pastilha de vidro. Antes da análise, as pastilhas de vidro foram dissolvidas em 100g de HNO₃ a 2% e diluídas gravimetricamente com mais HNO₃ a 2 % em frascos de polipropileno para obter um fator de diluição final de medição de 1:30.000. A análise por ICP-MS em solução foi realizada usando um procedimento semelhante ao descrito abaixo para as determinações de elementos traços.

Neste estudo, os materiais de referência W-2a (United States Geological Survey; USGS) e JP-1 (Serviço Geológico do Japão; GSJ) foram utilizados para calibração e controle de qualidade. Além disso, o material de referência AGV-2 (USGS) foi analisado como padrão de controle de qualidade secundário. Os resultados desse estudo reproduzem bem os valores sugeridos pelo GeoReM (Table S8) para esses materiais de referência.

2.5.2. *Elementos traços*

Para determinação de elementos traços, foram realizadas digestões em bomba fechada para garantir a digestão completa da amostra e a dissolução do espinélio. Uma alíquota de 100 mg de pó da amostra foi misturada com uma mistura de HF (~27 M) - HNO₃ (~14,5 M) na proporção de 4:1 e aquecida a 180 °C por quatro dias. A solução resultante foi evaporada a 100 °C e redissolvida em 3,5 ml de HCl a 6 M, seguida de aquecimento a 180 °C por um dia para garantir a dissolução completa dos fluoretos. A evaporação subsequente das soluções de amostra a 80 °C foi seguida pela remoção de qualquer excesso de Cl⁻ por evaporação repetida com 200 µl de HNO₃ a ~14,5 M. Os resíduos das amostras foram redissolvidos em 3 ml de HNO₃ a ~14,5 M e uma solução estoque com um fator de diluição nominal de ~1000 foi criada gravimetricamente. As soluções de estoque foram posteriormente diluídas para um fator de diluição gravimétrica nominal de 10.000, resultando em uma solução final de HNO₃ a 2 % pronta para a medição.

Seguindo o procedimento descrito em [Albut *et al.* \(2018\)](#), um padrão interno contendo ⁶Li, In, Re e Bi foi adicionado às amostras de calibração, materiais de referência de controle de

qualidade, amostras desconhecidas e amostras de monitoramento externo de deriva para corrigir a sensibilidade e a deriva instrumental. A calibração externa foi realizada usando o material de referência W-2a.

A precisão foi superior a 5 % para todos os elementos com base em análises repetidas dos materiais de referência AGV-2 e BHVO do USGS (Table S8). Para JP-1, material de referência de Peridotito altamente empobrecido do GSJ, a precisão foi melhor do 12 %. No entanto, exceções são encontradas para Nb (31 %), Ta (18 %) e U (55 %), (Table S8).

2.5.3. Razões isotópicas de Sr e Nd

A partir das soluções-padrão previamente preparadas para a análise de elementos traços, uma alíquota de cada amostra foi retirada e dissolvida em 1 ml de HCl 2,5 M para separação cromatográfica. Em resumo, o Sr e os elementos terras raras leves (LREE) foram separados usando resina de troca catiônica BioRad[®] AG 50W-X8 (200-400 mesh), seguido pela purificação do Sr usando resina Eichrom[®] Sr spec. A purificação subsequente do Nd foi alcançada por meio de um procedimento cromatográfico reverso utilizando resina Eichrom[®] Ln com HCl 0,5 M e 0,25 M.

Os separados de Sr foram impregnados com ativador de Ta em filamentos individuais de Re, e as medidas de razões isotópicas foram realizadas de forma dinâmica, utilizando um espectrômetro de massa por ionização termal Finnigan MAT262[®]. O fracionamento de massa analítico foi corrigido ao normalizar a razão $^{88}\text{Sr}/^{86}\text{Sr}$ de 8,375209, de acordo com a lei de decaimento exponencial. A medição repetida do padrão NBS SRM 987 resultou em uma razão $^{87}\text{Sr}/^{86}\text{Sr}$ de $0,710259 \pm 0,000025$ (2DP; n = 7; Table S5), que está em boa concordância com a reprodutibilidade de longo período de $0,710247 \pm 0,000033$ (2DP; n = 804).

Para a análise de isótopos de Nd, as amostras foram redissolvidas em HNO₃ 0,3 N e submetidas a medição utilizando um espectrômetro de massa acoplado indutivamente ThermoFisher Scientific NeptunePlus MC-ICP-MS operando em modo estático de baixa resolução. Para aumentar a sensibilidade, a introdução da amostra foi realizada com um nebulizador desolvatizante Aridus II, o que possibilitou a medição de soluções de Nd com 50 ng/g. O fracionamento de massa analítico foi corrigido usando a lei exponencial, com base na razão $^{146}\text{Nd}/^{144}\text{Nd}$ de 0,7219. Medições repetidas dos padrões La Jolla Nd e JNdi-1 resultaram em razões $^{143}\text{Nd}/^{144}\text{Nd}$ de $0,511830 \pm 0,000020$ (2DP, n = 4) e $0,512087 \pm 0,000018$ (2DP, n

= 11), respectivamente (Table S5). Esses valores estão condizentes com a reprodutibilidade de longo período de $0,511825 \pm 0,000011$ (2DP, n = 19) e $0,512075 \pm 0,000020$ (2DP, n = 47), respectivamente, obtidos pelo MC-ICP-MS. A precisão foi verificada por meio da digestão do material de referência padrão BHVO-2, o que resultou em uma razão $^{87}\text{Sr}/^{86}\text{Sr}$ de $0,703487 \pm 0,000010$ (2SE, n = 1) e uma razão $^{143}\text{Nd}/^{144}\text{Nd}$ de $0,512952 \pm 0,000009$ (2SE, n = 1), consistentes com os valores publicados ($^{87}\text{Sr}/^{86}\text{Sr} = 0,703465 \pm 0,000014$, 2DP, n = 3; $^{143}\text{Nd}/^{144}\text{Nd} = 0,512994 \pm 0,000017$, 2DP, n = 2; [Fourny et al., 2016](#)). Os brancos totais foram inferiores a 88 pg para Sr e 40,6 pg para Nd, sendo considerados insignificantes.

2.6. Isótopos de Os e elementos altamente siderófilos

As composições isotópicas de Os e os conteúdos dos PGE + Re foram analisados pelo método de diluição isotópica no Earth and Planets Laboratory (EPL), Carnegie Institution for Science, nos Estados Unidos. O procedimento envolve a digestão de pó de rocha total (1 g) usando 6 ml de uma solução de água regia reversa (2:1 $\text{HNO}_3\text{-HCl}$) em alta temperatura, aproximadamente 240 °C, em tubos de Carius de Pyrex por um período de 24 a 72 horas ([Shirey & Walker, 1995](#)). Durante esse processo, foram adicionados traçadores isotópicos mistos de ^{185}Re - ^{190}Os e ^{191}Ir - ^{99}Ru - ^{194}Pt - ^{105}Pd .

O Os foi isolado por meio de extração com solvente, utilizando CCl_4 de alta pureza e microdestilação em HBr, conforme procedimentos estabelecidos previamente (ex. [Reisberg & Meisel, 2002](#)). O Re, Ir, Pt e Pd foram separados por meio de cromatografia de troca de ânions utilizando resina AG1X8. As composições isotópicas de Os foram determinadas por espectrometria de massa por ionização termal negativa (N-TIMS) usando o equipamento Thermo-Fisher Triton. O Os foi analisado como OsO_3^- por meio de coleta de picos em um único coletor no multiplicador de elétrons.

As medições do padrão de Os EPL J-M, empregando o procedimento descrito acima, resultaram em um valor médio de $^{187}\text{Os}/^{188}\text{Os} = 0,17395 \pm 7$ (2σ). Logo, este valor está dentro da margem de erro da média de medições de alta precisão deste padrão usando copos de Faraday de $0,1739255 \pm 10$ (2σ). Os brancos procedimentais totais no EPL foram inferiores a 2 pg para Re, Os, Ir e Ru, 20 pg para Pd e 30 pg para Pt. As concentrações de Re, Ir, Ru, Pt e Pd foram determinadas usando o espectrômetro de massa Thermo Scientific iCAP Q ICP-MS.

2.7. Geoquímica de rocha total e idade K-Ar do basalto hospedeiro

A geoquímica de rocha total de três rochas basálticas (Y566, Y570, Y571) foi determinada utilizando as instalações do Instituto de Pesquisas de Terremotos da Universidade de Tóquio, Japão. As abundâncias de elementos maiores e traços selecionados (Sc, V, Cr, Co, Ni, Zn, Ga, Rb, Sr, Y, Zr, Nb e Ba) foram analisadas usando um espectrômetro de fluorescência de raios-X (XRF, PW2400; Philips Japan Ltd.), enquanto as abundâncias de outros elementos traços (Cs, terras raras, Ta, Hf, Pb, Th e U) foram analisadas com um ICP-MS (Plasma Quad 3; VG Scientia Holdings AB) conectado a um sistema de ablação a laser que utiliza um laser Nd:YAG de 213 µm com frequência quadruplicada (UP-213; New Wave Research Inc.). Tanto os dados de elementos maiores quanto de elementos traços foram obtidos por XRF e LA-ICP-MS, usando as mesmas pastilhas de vidro.

As idades K-Ar foram determinadas usando o método de sensibilidade não aditivada no Centro de Pesquisa Geoquímica, Escola de Pós-Graduação de Ciências da Universidade de Tóquio, Japão. As análises dos isótopos de Ar foram realizadas usando um VG-5400 modificado internamente (MS-III). Aproximadamente 0,5 g de amostras de rocha total em uma fração entre 60 e 80 mesh foram fundidas (~1700 °C) em alto vácuo. Com base em medidas repetidas do padrão atmosférico contendo $1,5 \times 10^{-7}$ cm³ STP de ⁴⁰Ar, as incertezas na sensibilidade de ⁴⁰Ar e nas razões ⁴⁰Ar/³⁶Ar são estimadas em 5% e 0,2%, respectivamente. A concentração de K foi determinada por XRF (veja detalhes acima). Os procedimentos aplicados para a datação K-Ar foram descritos em detalhes por [Nagao et al. \(1991\)](#).

CAPÍTULO 3

ARTIGO CIENTÍFICO

3. ARTIGO CIENTÍFICO

3.1. Manuscrito

Grenville-age highly-depleted spinel peridotites metasomatized by high-K hydrous alkaline melt in the Patagonian back-arc

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Abstract

Here we report the first investigation of a new suite of spinel-bearing mantle xenoliths from Canquel Plateau, Patagonia. They are highly-depleted Mesoproterozoic (1.3–0.9 Ga) harzburgites and clinopyroxene-poor lherzolites characterized by typical indicators of partial melt extraction, such as low whole-rock Al_2O_3 and CaO contents (both <1.5 wt.%), high Mg# of silicate phases (90–95), and Cr-rich spinel (Cr# 0.20–0.42). The depletion of incompatible highly siderophile elements (Re, Ru, Pd) relative to the primitive upper mantle reinforces high-degrees of melt extraction. The existence of phlogopite, a hydrated and K-rich mineral, coupled with abundant alkaline glass veins containing high- and low- SiO_2 compositions, are undoubtable evidence of modal metasomatism. The light rare earth element (LREE) enrichment observed in clinopyroxene, as well as the whole-rock U-shaped REE pattern are also evidence of metasomatism. The melt-rock interaction is corroborated by major (i.e., Mg# vs. basaltic elements) and trace (i.e., La/Yb_N vs. Sr/Y and Ti/Eu ; and Sr vs. Ti/Eu) element contents of pyroxenes. Moreover, the calculated compositions of melts in equilibrium with clinopyroxene coexisting with phlogopite, as well as whole-rock Sr-Nd isotopic ratios suggest interaction with subduction-related components. Remarkably, the metasomatic overprint is selectively prominent in the deeper (~70 km) mantle column compared to the shallower (~50 km). These features are consistent with the hybridization of the depleted lithosphere with a K-rich alkaline melt. Considering the long-lived subduction system established in the Patagonia back-arc throughout the Phanerozoic, it is most likely that the metasomatic event recorded by Los Gemelos mantle xenoliths is associated with the upwelling of asthenospheric material through a slab-window during the Paleocene to Eocene, which caused the partial melting of the oceanic crust.

Highlights

- Highly-depleted SCLM formed in Patagonia during the Rodinia consolidation in the Mesoproterozoic (1.3-0.9 Ga)
- Phlogopite was generated in the deeper mantle column through the percolation of a high-K hydrous alkaline melt
- This melt is probably derived from the partial melting of oceanic crust
- High-K alkaline glass veins were produced shortly before the mantle xenoliths were transported to the surface by the host magma in the Pleistocene
- These veins filled hydraulic fractures without significant reaction with the peridotitic paragenesis

Keywords: Spinel peridotites, Partial melting, Mesoproterozoic, Hydrous and high-K mantle metasomatism, Sr, Nd and Os isotopes, Patagonian mantle.

3.2. Introduction

Several subduction-related components, such as oceanic crust (generating silicate or CO₂-rich melts), sediments, and seawater have strong influence in the complex tectonic evolution recorded by the chemically and lithologically heterogeneous subcontinental lithospheric mantle (SCLM) beneath Patagonia (Stern *et al.*, 1999; Gorrington & Kay, 2000; Laurora *et al.*, 2001; Kilian & Stern, 2002; Rivalenti *et al.*, 2004, 2007; Schilling *et al.*, 2008, 2017; Ntaflos *et al.*, 2007; Bjerg *et al.*, 2005, 2009; Dantas *et al.*, 2009; Faccini *et al.*, 2013; Mundl *et al.*, 2015, 2016; Jalowitzki *et al.*, 2016, 2017; Melchiorre *et al.*, 2020; Novais-Rodrigues *et al.*, 2021; Bertotto *et al.*, 2021, 2022). Mantle metasomatism in this area has been mainly associated with the recycling of materials derived from the Andean subduction system along the western margin of South America during the Phanerozoic. The interactions between peridotite and melt/fluids derived from subducted slabs can produce modal metasomatism that is related to the formations of mica (phlogopite) and/or amphibole (pargasite), which are exotic and hydrated mineral phases in the typical anhydrous mantle assemblage composed of olivine, orthopyroxene and clinopyroxene plus an aluminum phase (usually spinel, but also

garnet or plagioclase). Moreover, glass veins containing variable compositions and secondary minerals resulting from reactions between melt-rock are also evidence of mantle metasomatism. This process is normally associated with the chemical enrichment of the coexisting primary mineral phases, being characterized by high incompatible over less incompatible element ratios (i.e., light rare earth elements - LREE over heavy rare earth elements - HREE). When secondary minerals are not formed, this process is known as “cryptic metasomatism”. However, in Patagonia, older vestiges of mantle metasomatism are scarce, especially due to the replacement of these compositional signatures (depletion or even further enrichment) by younger events related to modern (Cenozoic) subduction zones ([Melchiorre et al., 2020](#)).

Aiming to constrain the origin and time of both depletion and metasomatic events recorded in the SCLM beneath the northern edge of the Deseado Massif, we report the occurrence of a new suite of mantle xenoliths hosted by a Pleistocene (0.54–0.34 Ma; Table S1) alkaline lava from Los Gemelos, a twin post-plateau monogenetic cone located in the Canquel Plateau (or Meseta del Canquel). These mantle-derived xenoliths provide a valuable opportunity to investigate the tectono-magmatic processes involved in the long-lived and complex evolution of the Patagonian back-arc. To achieve this objective, we present mineral and whole-rock geochemistry, Sr-Nd-Os isotopic composition, and highly siderophile element abundances (HSE; platinum group elements - PGEs + Re) for 16 spinel-bearing xenoliths.

3.3. Geological setting

The Canquel Plateau represents the austral prolongation of the Somuncura large igneous province, being characterized by several superimposed Cenozoic lava flows covering an area of ca. 1800 km² ([Anselmi et al., 2004](#); [Massaferro et al., 2017](#)) (Figure 1). These flows are essentially composed of olivine basalts from the Sierra Cuadrada Formation, which records the early stages of volcanic activity in the Oligocene but have predominant Miocene ages (23–21 Ma) ([Anselmi et al., 2004](#) and references therein). However, younger (6.20–2.26 Ma) and less voluminous post-plateau spatter cones were reported at the top of the Canquel Plateau ([Anselmi et al., 2004](#); [Massaferro et al., 2014](#)). In this context, the Pleistocene (0.54–0.34 Ma; Table S1) lava flow from the Los Gemelos monogenetic volcanoes host the mantle xenoliths investigated here (Figure 1). The large volume of magma generated essentially during the

Miocene in the Somuncura plateau, Canquel plateau, and Paso de Indios region has been ascribed to the existence of a “wet” asthenospheric upwelling caused by dehydration-induced melting of mantle transition zone in Central-North Patagonia ([Honda *et al.*, 2006](#); [Orihashi *et al.*, 2006, 2010](#); [Navarrete *et al.*, 2020](#)).

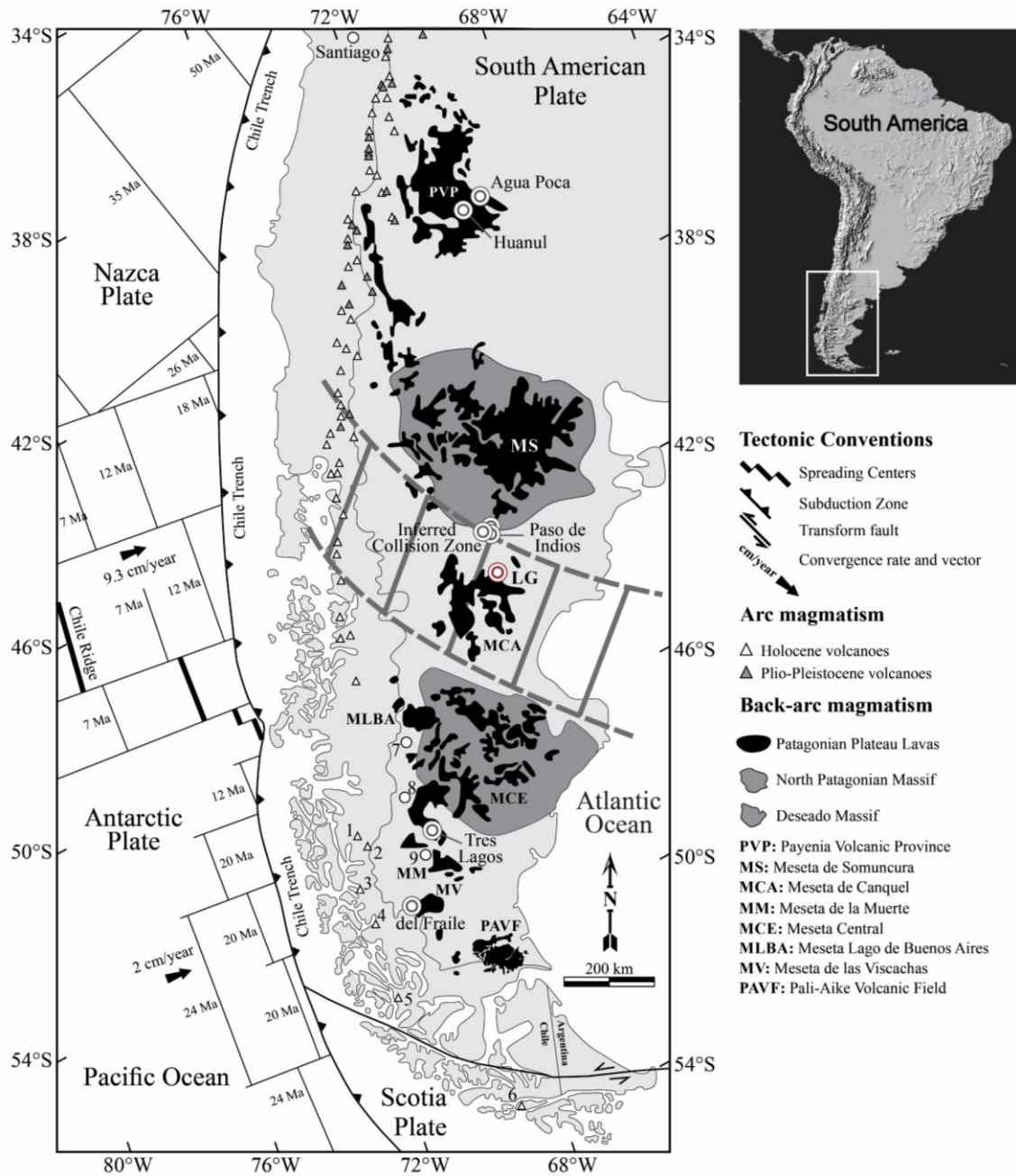


Figure 1. Map of southern South America showing the location of the Los Gemelos volcano (LG). Locations of depleted mantle xenoliths are from Agua Poca, Huanul, and Tres Lagos while metasomatized xenolith samples (by adakite melts) are from Paso de Indios region (Chenque, Matilde, and León) and Cerro del Fraile. The occurrence of adakite lavas in the Austral Volcanic Zone (AVZ) is indicated as 1 = Lautaro, 2 = Viedma, 3 = Aguilera, 4 = Reclus, 5 = Mt. Burney, 6 = Cook Island, whereas in the back-arc region of Southern Volcanic Zone is 7 = Cerro Pampa, 8 = Chaitén, and 9 = Puesto Nuevo.

The basement rocks of this region are covered by the lavas from Canquel Plateau, as well as by the thick Mesozoic to Cenozoic sedimentary successions of the rift-related Golfo San Jorge basin (Renda *et al.*, 2019). Los Gemelos volcanoes outcrop in the inferred collision

zone developed during the Carboniferous-Permian between the Deseado and the North Patagonian massifs (Pankhurst *et al.*, 2006; Schilling *et al.*, 2017). According to the tectonic model proposed by these authors, the SCLM beneath Los Gemelos belongs to the northern edge of the Deseado Massif.

The basement ages estimated for these two massifs suggest the occurrence of two main episodes of crust formation. The oldest event is evidenced by rocks with Paleo- to Mesoproterozoic model ages and relatively abundant detrital zircon populations related to the Rodinia consolidation (1300–900 Ma), whereas the youngest rocks have Pampean and Brasiliano ages related to Rodinia fragmentation and to the early stages of Gondwana consolidation (750–550 Ma) (Pankhurst & Rapela, 1995; Pankhurst *et al.*, 2003, 2006, 2014; Hervé *et al.*, 2003; Ramos, 2010; Martínez Dopico *et al.*, 2011; Schilling *et al.*, 2017). Therefore, Patagonia appears to have integrated the Rodinia supercontinent during the Mesoproterozoic, being detached during its break-up, and in the Neoproterozoic being accreted to Gondwana (Ramos, 2010).

In the upper Paleozoic, the collision of the Deseado Massif against the North Patagonian Massif caused the rejuvenation, erosion, and thinning of the SCLM in the south-southwestern margin of the North Patagonian Massif, together with a slab break-off and asthenospheric upwelling (Pankhurst *et al.*, 2006; Schilling *et al.*, 2017). In addition, in the Cenozoic, two other slab-windows were opening in the west margin of Patagonia, acting as potential mechanisms to metasomatize the SCLM (e.g., Cande & Leslie, 1986; Breitsprecher & Thorkelson, 2009; Jalowitzki *et al.*, 2017). During the Paleocene to Eocene, the Farallon and Aluk oceanic plates collided against the South American continental plate. The triple junction formed through their intersection with the oceanic trench migrated southward along the western margin of South America. To the south, the continuous subduction of the Chile Ridge spreading center has been migrating from the southern edge of Patagonia to the latitude of the Taitao Peninsula (~46°S), forming a trench-ridge-trench triple junction for the last 16 Ma (e.g., Cande & Leslie, 1986; Breitsprecher & Thorkelson, 2009).

3.4. Analytical methods

In this study, we selected 16 mantle xenoliths large enough for geochemical and isotopic analysis that were also examined in thin sections. The samples were firstly

disaggregated by the SELFRAG™ Lab system at the University de Brasília (UnB). Subsequently, only weathered crystals were removed from the disaggregated sample by carefully handpicked under a binocular microscope. Mineral major element analyses were determined in the 16 selected samples (Table S2), whereas mineral trace and rare earth element (REE) abundances were obtained in a subset of 10 representative samples (Table S3). Additionally, sample aliquots (obtained after SELFRAG™ separation and the removal of weathered minerals) were finely ground into powder using an agate mortar. They were analyzed for their whole-rock major and trace element compositions ($n = 9$, Table S4), Sr-Nd isotopic ratios ($n = 7$, Table S5), Os isotopic compositions and HSE abundances ($n = 6$) (Table S6). In addition, the whole-rock geochemistry and K-Ar age of three volcanic rocks hosting the mantle xenoliths from Los Gemelos were determined (Table S1).

3.4.1. *Electron Probe Microanalysis*

Major elements of mineral phases (olivine, orthopyroxene, clinopyroxene, spinel, and phlogopite) and well-developed glass veins from Los Gemelos mantle xenoliths were determined by wavelength-dispersive spectrometry (WDS) using a JEOL JXA-8230 Electron Probe Micro Analyzer (EPMA) at the Laboratório de Microsonda Eletrônica, Instituto de Geociências (IG), UnB, Brazil. The analyses were done using an accelerating voltage of 15 kV, a beam current of 10 nA, and a beam diameter of 1 μm in the spot mode. For phlogopite measurements, we employed the circle mode with 5 μm . Counting times on peak were 10 s per analyte peak and 5 s for the background. All results were reduced using an in-house ZAF correction program. The measurements were calibrated with the following mineral standards: andradite (CaO and FeO), albite (Na₂O), forsterite (MgO), microcline (K₂O, Al₂O₃ and SiO₂), pyrophanite (TiO₂ and MnO), chromium oxide (Cr₂O₃), nickel oxide (NiO), topaz (F), and vanadinite (V₂O₃ and Cl).

3.4.2. *Laser ablation inductively coupled plasma mass spectrometry*

Clinopyroxene and phlogopite trace elements were determined at the Laboratory of the Centro Interdipartimentale Grandi Strumenti (CIGS), the Università di Modena e Reggio Emilia, Italy. A Thermo Fisher Scientific XSERIES-2 mass spectrometer coupled to a New

Wave UP 213 laser ablation was used. Data reduction was performed with the Thermo Fisher Scientific PlasmaLab® software using NIST610 (SD <4.0 %) (Jochum *et al.*, 2005) and BCR-2 (SD <3.8 %) (Jochum *et al.*, 2015) as external standards (Table S7). The analysis conditions were laser spot 55 µm, output 70 % wt., Rep. Rate 10 Hz, dwell time 30 s and laser beam fluency at 8 J/cm². The sample surface was first ablated with 65 µm spot size, 45 % wt., 10 Hz repetition rate. The isotope ⁴⁴Ca was used as internal standard for clinopyroxene, whereas isotope ²⁹Si was used for phlogopite.

3.4.3. Whole-rock geochemistry and radiogenic Sr-Nd isotopes

Sample preparation, whole-rock geochemistry, and radiogenic Sr-Nd isotope analyses were done at the Isotope Geochemistry laboratories of the University of Tuebingen, Germany. Loss on ignition (LOI) was determined by placing an aliquot of 100 mg of sample in a furnace at 1050 °C for one hour, cooling in a desiccator, and reweighing. Due to the limited sample material, a ThermoFisher Scientific iCAP-Qc ICP-MS was used in this study to determine major and trace element compositions.

3.4.3.1. Major elements

For major element analysis, 100 mg of material was mixed with 500 mg of Merck® Spectromelt® A 12 (mixture of 66 % di-lithium tetraborate and 34 % lithium metaborate) and fused at 1200 °C. The resulting glass beads were dissolved in 2 % HNO₃, gravimetrically diluted, and analyzed using a similar procedure described below for trace element determinations. Reference materials W-2a (United States Geological Survey; USGS), JP-1 (Geological Survey of Japan; GSJ), and AGV-2 (USGS) were used for calibration and quality control. Our results agree well with the GeoReM preferred values (Table S8).

3.4.3.2. Trace elements

For trace element determinations, closed bomb digestions were carried out to ensure complete sample digestion and dissolution of spinel. A 100 mg sample powder was mixed with a 4:1 mixture of HF (~27M) - HNO₃ (~14.5M) and heated at 180 °C for four days. The resulting solution was evaporated at 100 °C and re-dissolved in 3.5 mL 6M HCl, followed by heating at 180 °C for one day to ensure the destruction of fluorites. Subsequent evaporation

was followed by removing any excess of Cl⁻ by repeated evaporation to incipient dryness with ~14.5M HNO₃. The sample residues were re-dissolved in 3 mL 14.5M HNO₃, and a stock solution with a nominal dilution factor of ~1000 was created gravimetrically. The stock solutions were further diluted to a nominal gravimetric dilution factor of 10000, yielding a final 2 % HNO₃ solution ready for measurement.

Following the procedure described in detail by [Albut *et al.* \(2018\)](#), an internal standard carrying ⁶Li, In, Re and Bi was added to calibration standards, quality control reference materials, sample unknowns, and external drift monitor samples to account for sensitivity and instrumental drift. External calibration was performed using the averaged intensities at three different dilution factors of reference material W-2a.

Precision was better than ca. 5 % for all elements based on repeated analyses of USGS reference materials AGV-2 and BHVO (Table S8). Based on JP-1, a highly depleted peridotite reference material from GSJ, precision was better than 12 %, except for Nb (31 %), Ta (18 %), and U (55 %) (Table S8).

3.4.3.3. Sr and Nd isotopic ratios

Between 100 to 400 mg of sample powders were weighed into Savillex Teflon beakers and digested via HF (48%) – HNO₃ (65%) acid attack at 100 °C. Digested samples were dried and transformed into chlorides in 6 N HCL. The samples were redissolved in 1 ml of 2.5M HCl for chromatographic separation. In brief, Sr and light (L) REE were separated using BioRad[®] AG 50W-X8 cation exchange resin (200-400 mesh), followed by Sr purification using Eichrom[®] Sr spec resin. Subsequent purification of Nd was achieved through a reverse ion chromatographic procedure using Eichrom[®] Ln resin with 0.5M and 0.25M HCl.

Sr separates were loaded with Ta-activator on Re single filaments, and isotope ratio measurements were performed in dynamic mode using a Finnigan MAT262[®] thermal ionization mass spectrometer (TIMS). Analytical mass fractionation was corrected with an ⁸⁸Sr/⁸⁶Sr ratio of 8.375209 and exponential law. Repeated measurement of NBS SRM 987 resulted in an ⁸⁷Sr/⁸⁶Sr ratio of 0.710259 ± 0.000025 (2SD; n = 7; Table S5), which is in good agreement with the long-term reproducibility of 0.710247 ± 0.000033 (2SD; n = 804).

For Nd isotope analysis, samples were re-dissolved in 0.3N HNO₃ and were measured using a ThermoFisher Scientific NeptunePlus MC-ICP-MS in low-resolution static mode. To enhance sensitivity, an Aridus II desolvating nebulizer was employed for sample introduction,

allowing measurement of 50 ng/g Nd solutions. Analytical mass fractionation was corrected using exponential law with a $^{146}\text{Nd}/^{144}\text{Nd}$ ratio of 0.7219. Repetitive measurements of the La Jolla Nd and JNdi-1 standards yielded $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.511830 ± 0.000020 (2SD, $n = 4$) and 0.512087 ± 0.000018 (2SD, $n = 11$), respectively (Table S5). These values are in good agreement with the long-term reproducibility of 0.511825 ± 0.000011 (2SD, $n = 19$) and 0.512075 ± 0.000020 (2SD, $n = 47$), respectively, obtained by the MC-ICP-MS. Accuracy was checked through digestion of the rock reference standard material BHVO-2, which gave an $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.703487 ± 0.000010 (2SE, $n = 1$) and $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.512952 ± 0.000009 (2SE, $n = 1$), consistent with published values ($^{87}\text{Sr}/^{86}\text{Sr} = 0.703465 \pm 0.000014$, 2SD, $n = 3$; $^{143}\text{Nd}/^{144}\text{Nd} = 0.512994 \pm 0.000017$, 2SD, $n = 2$; Fourny *et al.*, 2016). Total procedural blanks were below 88 pg for Sr and 40.6 pg for Nd, thus considered negligible.

3.4.4. *Re-Os and highly siderophile elements*

Re-Os isotopic compositions and HSE contents were determined using the isotope dilution method at the Earth and Planets Laboratory (EPL), Carnegie Institution for Science, United States of America. Whole-rock powder (1 g) was digested using 6 ml of reverse aqua regia (2:1 $\text{HNO}_3\text{-HCl}$) at high-temperature ($\sim 240^\circ\text{C}$) in Pyrex Carius tubes over a period of 24–72 h (Shirey & Walker, 1995), together with mixed spikes of $^{185}\text{Re}\text{-}^{190}\text{Os}$ and $^{191}\text{Ir}\text{-}^{99}\text{Ru}\text{-}^{194}\text{Pt}\text{-}^{105}\text{Pd}$. Osmium was separated by solvent extraction employing high-purity CCl_4 and microdistillation in HBr following established techniques (e.g., Reisberg & Meisel, 2002). Rhenium, Ir, Pt and Pd were separated via anion exchange chromatography using AG1X8 resin. Osmium isotopic compositions were determined by negative thermal ionization mass spectrometry (N-TIMS) using the Thermo-Fisher Triton. Osmium was analyzed as OsO_3^- by single-collector peak-hopping in the electron multiplier. Measurements of the EPL J-M Os standard using this procedure during this study gave an average value of $^{187}\text{Os}/^{188}\text{Os} = 0.17395 \pm 7$ (2SD). This value is within error of the mean of high-precision measurements of this standard using faraday cups of 0.1739347 ± 20 (2SD). Total procedural blanks at EPL were less than 2 pg for Re, Os, Ir and Ru, 20 pg for Pd, and 30 pg for Pt. Rhenium, Ir, Ru, Pt and Pd isotopic compositions for isotope dilution concentration determinations were measured using the Thermo Scientific iCAP Q ICP-MS.

3.4.5. *Host lava whole-rock geochemistry and K-Ar age*

The whole-rock geochemistry of three basaltic rocks (Y566, Y570, Y571) was determined using the facilities of the Earthquake Research Institute, University of Tokyo, Japan. Major and selected trace element abundances (Sc, V, Cr, Co, Ni, Zn, Ga, Rb, Sr, Y, Zr, Nb, and Ba) were analyzed using X-ray fluorescence spectrometer (XRF, PW2400; Philips Japan Ltd.), whereas the abundances of other trace elements (Cs, REEs, Ta, Hf, Pb, Th, and U) were analyzed with an ICP-MS (Plasma Quad 3; VG Scienta Holdings AB) connected to a laser ablation system using a frequency-quadrupled 213 μm Nd: YAG laser (UP-213; New Wave Research Inc.). Both major element and trace element data were obtained by XRF and LA-ICP-MS using the same glass beads.

The K-Ar ages were determined using the unspiked sensitivity method at the Geochemical Research Center, Graduate School of Science, University of Tokyo, Japan. The Ar isotope analyses were performed using an in-house modified VG-5400 (MS-III). Approximately 0.5 g of whole-rock samples in a fraction between 60 and 80 mesh were melted (~ 1700 °C) in a high vacuum. Based on repeated measurements of the atmospheric standard containing 1.5×10^{-7} cm³ STP of ⁴⁰Ar, the uncertainties on ⁴⁰Ar sensitivity and ⁴⁰Ar/³⁶Ar ratios are estimated as 5 % and 0.2 %, respectively. The K concentration was determined by XRF (see details above). The procedures applied for K-Ar dating were described in detail by [Nagao et al. \(1991\)](#).

3.5. RESULTS

3.5.1. *Petrography*

The 16 xenoliths from Los Gemelos are fresh peridotites ~ 4 –12 cm in size. Modal proportions of mineral phases were determined by point-counting using the JMicroVision software ([Roduit, 2007](#)). For each thin section (24 x 40 mm), between 2034 and 3826 points were counted. Based on modal analysis, spinel peridotites were classified as anhydrous clinopyroxene-poor lherzolites (LG-01 and LG-10) and as hydrated- (phlogopite-bearing LG-08 and LG-12) and anhydrous-harzburgites (other samples) (Figure S1 and Table S9). Olivine (59–84 vol.%) and orthopyroxene (10–36 vol.%) are the two dominant phases in all samples,

whereas clinopyroxene (2–9 vol.%), spinel (1–5 vol.%), and phlogopite (<1 vol.%) occur in lower modal proportions (Table S9).

Peridotites from Los Gemelos are texturally equilibrated as indicated by triple point junctions (120°) and curvilinear grain boundaries (Figure 2a). They display little deformation features with textures varying from medium to coarse protogranular, which is associated with the harzburgitic paragenesis mainly as a product of partial melting ([Mercier & Nicolas, 1975](#)) (Figure 2a). Crystal sizes are in the range of 0.3 and 7.2 mm. The contacts between the xenoliths and the host alkaline-basalt are always straight with no evidence of basalt-peridotite interactions.

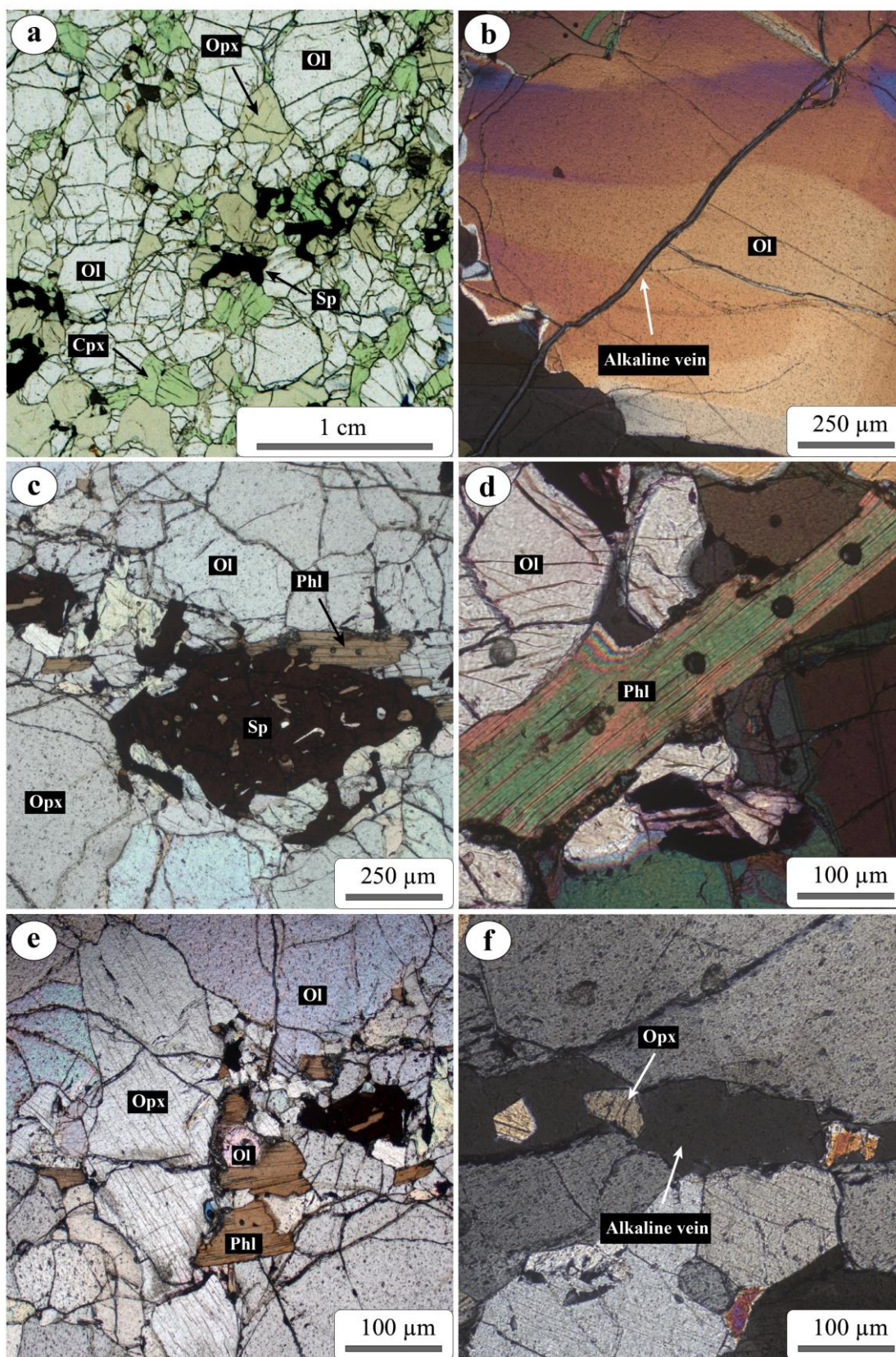


Figure 2. Representative images showing the main mineralogical and textural features observed in mantle xenoliths from Los Gemelos. **(a)** Detail of a scanned image from a thin section showing the representative mineral assemblage composed by olivine (Ol), orthopyroxene (Opx), clinopyroxene (Cpx), and spinel (Sp) (sample LG-10). **(b)** Photomicrograph (crossed-polarized light) of a well-developed vein of adakite composition crosscutting an olivine crystal (sample LG-06). **(c)** Photomicrograph highlighting the occurrence of phlogopite crystals as inclusions and around spinel grains (sample LG-12). **(d)** Photomicrograph (crossed-polarized light) showing an example of a euhedral phlogopite crystal in which the hole trail represents the laser ablation in the ICP-MS (sample LG-08). **(e)** Photomicrograph showing the localized occurrence of phlogopite as euhedral crystals and as inclusion in spinel (sample LG-12). **(f)** Photomicrograph (crossed-polarized light) showing the crystallization of reaction minerals (type 2) within adakite veins (sample LG-02).

Olivine is generally subhedral and displays variable sizes (0.6–5.7 mm). Large olivine crystals often exhibit fractures and kink-bands (Figure 2b) and, occasionally, contain inclusions of spinel. *Orthopyroxene* is subhedral to anhedral and displays larger sizes with respect to other phases (3.3–7.2 mm) (Figure 2a). In some samples, clinopyroxene exsolution lamellae are observed along crystallographic controlled directions. *Clinopyroxene* is smaller than olivine and orthopyroxene (0.3–2.3 mm), commonly occurring as anhedral crystals characterized by intense green color in textural equilibrium with the other coexisting minerals (Figure 2a). Clinopyroxene crystals occur either as discrete grains or in clusters. *Spinel*s vary in color from brown to reddish-brown. They usually appear as interstitial minerals and seem to be late textural phases, ranging from amoeboid to vermicular shapes with sizes varying from 0.1 to 1.7 mm (Figure 2a, c). *Phlogopite* has sizes in the range of 0.4 and 1.3 mm and occurs as tabular isolated crystals or associated with spinel (in contact), sometimes as inclusions (Figure 2c-e and Figure 3a-d). The isolated phlogopite is larger, euhedral, and texturally equilibrated with peridotite assemblages (Figure 2d-e and Figure 3**Erro! Fonte de referência não encontrada.**c-d). It is not presente in kelyphitic selvages, serpentized veins, or as overgrowths around other minerals, and thus can be defined as primary phlogopite in the sense of [Erlank *et al.* \(1987\)](#). Conversely, phlogopite around and as inclusions in spinel crystals are smaller and subhedral (Figure 2c, e and Figure 3a-b). *Metasomatic veins* with hydraulic fracturing and trails of secondary inclusions of olivine, orthopyroxene, and clinopyroxene are observed in samples LG-02, LG-06, LG-08, LG-12, LG-14, and LG-15 (Figure 2b, f and Figure 3e-f). These veins do not occur surrounding the mineral boundaries, that is, they are not an intergranular material.

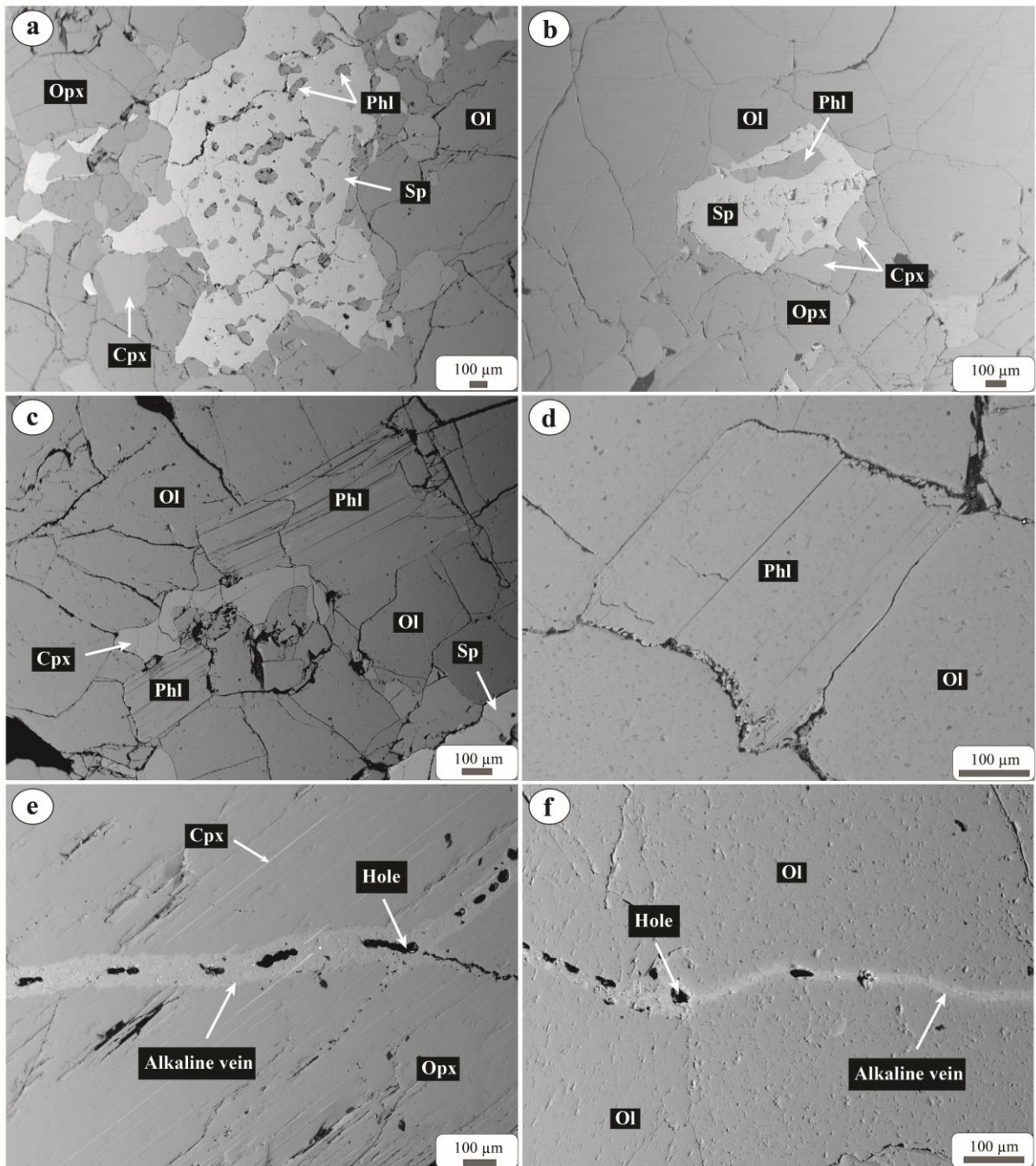


Figure 3. Representative backscattered electron images (BSE) of glass veins with adakite composition and phlogopites (Phl), which define modal metasomatism. **(a-b)** Spinel (Sp) crystals containing abundant silicate inclusions, as phlogopite (sample LG-08). **(c-d)** Well-crystallized euhedral phlogopite crystals (sample LG-08). **(e-f)** Adakite glass veins with variable thicknesses crosscutting, respectively, orthopyroxene (Opx) (sample LG-02) and olivine (Ol) (sample LG-06). Note that Opx has clinopyroxene exsolution lamellae. The black areas observed within adakite veins indicate holes produced by polishing of the thin section.

3.5.2. Mineral chemistry

Individual measurements and the representative (average) compositions of olivine, orthopyroxene, clinopyroxene, spinel, phlogopite, and veins containing alkaline composition are given in Table S2. Negligible core-rim variation ($<0.5\%$) between the main oxides and alkaline veins was observed. For this reason, the average compositions were employed in the major element diagrams (Figure 4 and Figure S3). Trace and rare earth element concentrations were determined only for clinopyroxene and phlogopite (Table S3).

Olivine ($\text{Fo}_{0.90-0.91}$) is characterized by high Mg# (0.90–0.92; $\text{Mg\#} = \text{Mg}/(\text{Fe}+\text{Mg})$ molar), and NiO content (0.37–0.43 wt.%), coupled with low CaO content (<0.04 wt.%).

Chromium spinel contains high Cr# (0.20–0.42; $\text{Cr\#} = \text{Cr}/(\text{Cr}+\text{Al})$ molar) and Mg# between 0.67 and 0.80, defining a negative correlation between these notations (Figure S2). The only sample showing variation in spinel composition is LG-02 which is expressed by two different Al_2O_3 (43.45 and 37.87 wt.%) and Cr_2O_3 (25.44 and 31.35 wt.%) contents, implying variable Cr# (0.28 and 0.36) (Figure S2).

Orthopyroxene is enstatite ($\text{En}_{0.90-0.91} \text{Fs}_{0.08-0.09} \text{Wo}_{0.07-0.10}$) with low Al_2O_3 (1.83–2.97 wt.%), TiO_2 (<0.08 wt.%) and CaO (0.34–0.53 wt.%) contents at a given high Mg# (0.91–0.92) (Figure S3).

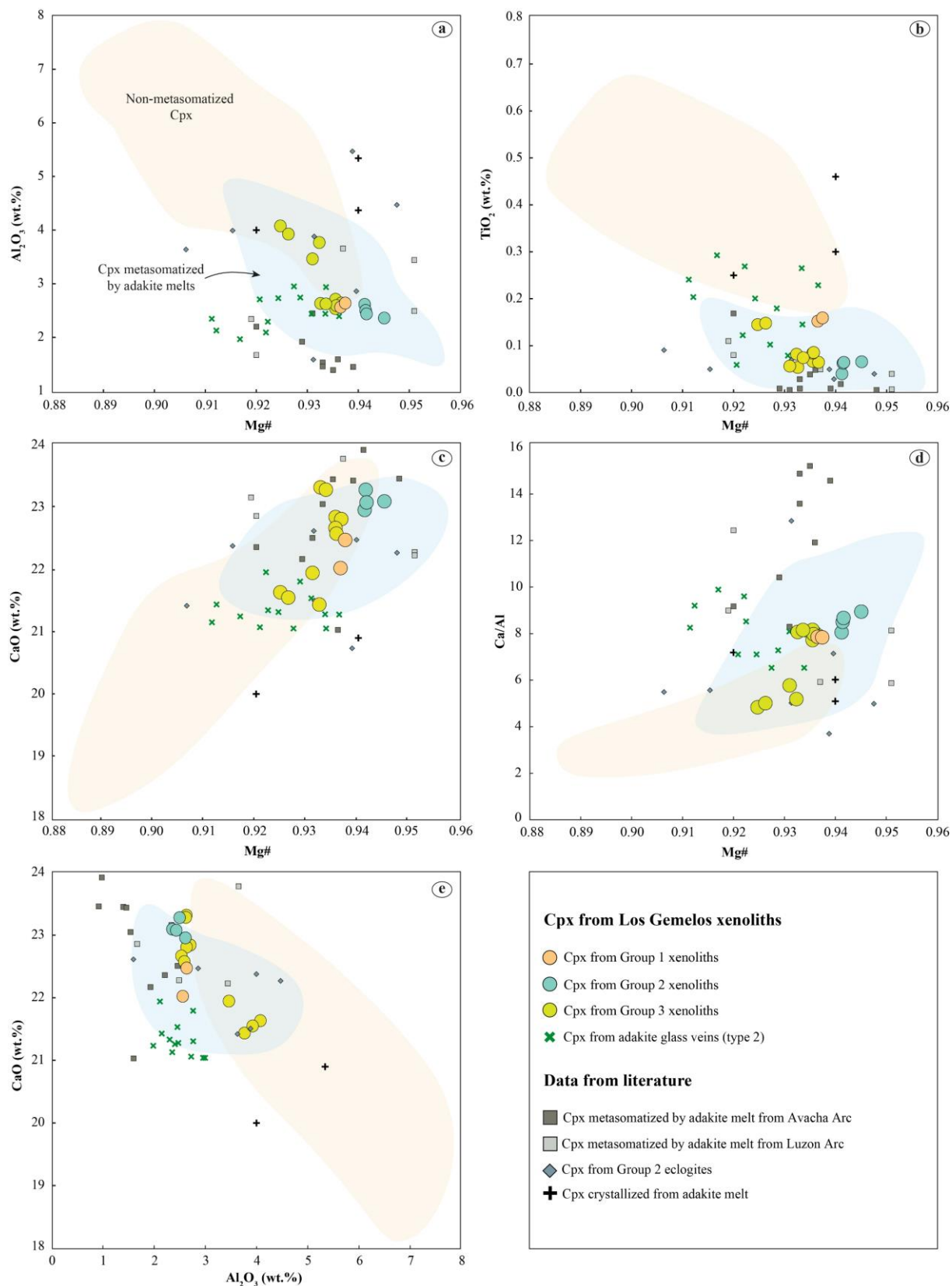


Figure 4. Selected major elements (Al_2O_3 , TiO_2 , CaO , and Ca/Al) vs. $\text{Mg\#}$ (a-d) and Al_2O_3 vs. CaO (e) of clinopyroxenes (Cpx) classified as primary (type 1) and crystallized within adakite glass veins (type 2) from Los Gemelos. Our samples show compositional affinity with clinopyroxenes metasomatized by adakite melts (light blue field) of mantle xenoliths from Cerro de los Fraile and Paso de Indios region (Faccini *et al.*, 2013; Bertotto *et al.*, 2021). The field defined for non-metasomatized clinopyroxenes from Patagonia (light yellow) includes mantle xenoliths from Agua Poca, Huanul, and Tres Lagos (Ntaflos *et al.*, 2007; Bertotto *et al.*, 2013, 2022). For comparison, we plotted clinopyroxenes metasomatized by adakite melts from Avacha and Luzon arcs (Arai *et al.*, 2003, 2004; Ishimaru *et al.*, 2007), from Group 2 eclogites (Viljoen *et al.*, 2005), and experimentally crystallized by adakite melts (Corgne *et al.*, 2018).

Clinopyroxene is a high- $\text{Mg\#}$ (0.92–0.95) Cr-diopside endmember ($\text{Cr}_2\text{O}_3 = 0.72\text{--}1.71$ wt.%; $\text{En}_{0.47\text{--}0.50}$ $\text{Fs}_{0.03\text{--}0.04}$ $\text{Wo}_{0.47\text{--}0.49}$) with intermediate to high Ca/Al (molar) ratios (4.83–8.94) in which Al_2O_3 varies from 2.36 to 4.08 wt.% and CaO between 21.44 and 23.32 wt.%, whereas TiO_2 is <0.16 wt.% (Figure 4). CaO contents and Ca/Al ratios show positive correlations with $\text{Mg\#}$, whereas Al_2O_3 and TiO_2 display negative correlations with $\text{Mg\#}$ (Figure 4). A negative correlation also is observed in CaO vs. Al_2O_3 contents (Figure 4). Note that the increasing of $\text{Mg\#}$ is accompanied by depletion in Al_2O_3 and TiO_2 contents in both pyroxenes (Figure 4 and Figure S3). Based on the primitive mantle (PM, Sun & McDonough, 1989) normalized trace and REE diagrams (Figure 5), as well as on the petrographic observations (presence or absence of phlogopite and glass veins), clinopyroxenes from Los Gemelos peridotites are classified into three groups. clinopyroxenes from Los Gemelos peridotites are classified into three groups. Group 1 (G1) is comprised by xenoliths containing phlogopite (LG-08 and LG-12) and they show fractionated LREE over HREE pattern (i.e., $\text{Ce/Yb}_N = 8.79\text{--}10.11$, where $N = \text{PM}$ normalized values) (Figure 5a). In contrast, Group 2 (G2: LG-02, LG-06, LG-14, LG-15) and especially Group 3 (G3: LG-03, LG-09, LG-11, LG-16) contain spoon-like REE patterns defined by depletion from HREE to Pr (i.e., $\text{Pr/Yb}_N = 0.01\text{--}0.25$) with selective enrichment from Pr to LREE (i.e., $\text{La/Pr}_N = 1.50\text{--}6.89$) (Figure 5a). These two groups show Ce/Yb_N ratios lower than 0.32. Veins with alkaline Affinity were only observed in xenoliths from G1 and G2. In the PM-normalized trace element diagram, clinopyroxenes have positive anomalies of W, U, La, Pb, Mo (except for G1 xenoliths), and Eu, with negative anomalies of Rb, Ba, Th, K, Zr, Hf, and Ti (Figure 5b). Niobium and Ta negative anomalies are identified in samples from G1 and G2 (especially in the former), but G3 do not show clear depletion in these elements (Figure 5b). Although there are no clinopyroxene REE data for some xenoliths (LG-01, LG-04, LG-05, LG-07, LG-10, LG-13), they were included in G3 according with their clinopyroxene major element distribution (Figure 4), and lack of glass veins and phlogopite.

Phlogopite is restricted to xenoliths LG-08 and LG-12, showing homogeneous composition among euhedral crystals, as inclusions in spinel, or even as crystals in contact with spinel (e.g., $\text{Al}_2\text{O}_3 \sim 15$, $\text{K}_2\text{O} \sim 9$, $\text{MgO} \sim 22$, $\text{TiO}_2 \sim 2$, and $\text{FeO} \sim 3$ wt.%; Table S2). They show strong enrichment of LREE over HREE ($\text{Ce}/\text{Yb}_\text{N} = 15.32\text{--}22.08$) and pronounced positive Eu anomaly ($\text{Eu}/\text{Eu}^* = 4.13\text{--}10.88$, where $\text{Eu}/\text{Eu}^* = \text{Eu}_\text{N}/(\text{Sm}_\text{N} \times \text{Gd}_\text{N} \text{ or } \text{Tb}_\text{N})^{1/2}$) (Figure 6a). In the PM-normalized trace element diagram, these phlogopites are enriched in incompatible elements (i.e., Cs, Rb, Ba, K, Pb, Mo, Sr, and Li), and in high field strength field elements (HFSE, Nb-Ta-Ti) compared to REEs and Th (Figure 6b).

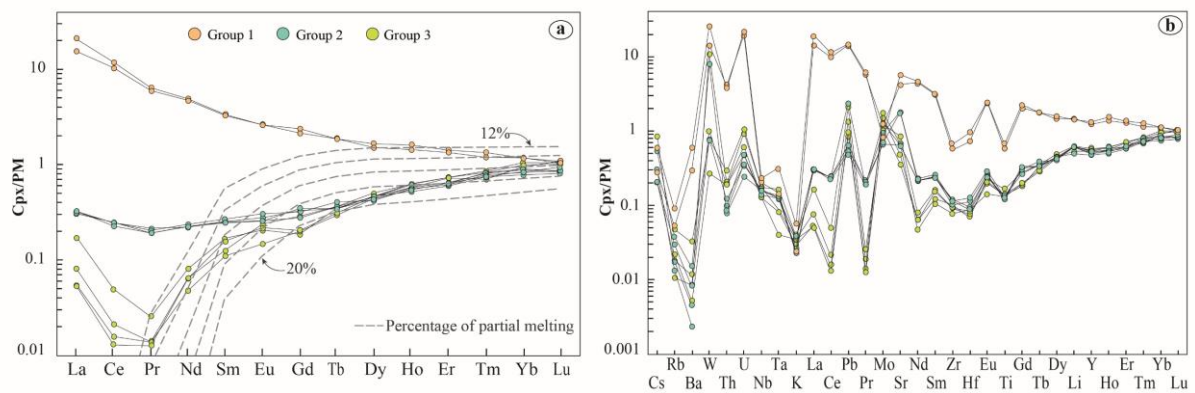


Figure 5. Rare earth element (a) and multielement (b) diagrams of clinopyroxenes (Cpx) from Los Gemelos. The concentrations are normalized to the primitive mantle (PM, [Sun & McDonough, 1989](#)). The partial melting estimates from 12 to 20 % are indicated by dashed lines with the increment of 2 % in (a). Details about the calculations are in the main text.

Well-developed *glass veins* (Figure 2b, f and Figure 3e-f) with alkaline compositions are widespread in spinel peridotites from G1 and G2. They have bimodal composition marked by melts with high- and low- SiO_2 contents (Figure S4 and Figure S5). High- SiO_2 melts have compositional affinity with typical adakites as demonstrated by their high SiO_2 (60.38–61.29, adakite >56 wt.%) Al_2O_3 (15.78–16.52 wt.%, adakite >15 wt.%) and Na_2O (4.40–4.54 wt.%, adakite >3 wt.%) coupled with low MgO (2.26–2.37 wt.%, adakite <3 wt.%) ([Defant & Drummond, 1990](#); [Martin et al., 2005](#); [Castillo, 2012](#) and references therein) (Figure S4a-c and Table S2). On the other hand, low- SiO_2 melts show low SiO_2 (48.75–50.93) coupled with high Al_2O_3 (16.56–17.62 wt.%), Na_2O (4.14–5.32 wt.%) and MgO (4.43–4.85 wt.%) (Figure S4a-c and Table S2). Both types of melts have high K_2O (2.84–4.37 wt.%, Figure S4d) and high alkali contents ($\text{K}_2\text{O} + \text{Na}_2\text{O}$ range between 7.37–9.59 wt.%) compared to adakite lavas from Patagonia ($\text{K}_2\text{O} < 2.7$ wt.% and alkalis <7.0 wt.%; [Stern & Kilian, 1996](#); [Mahlburg Kay](#)

et al., 1993; Ramos *et al.*, 2004; Table S2). The high K₂O contents of these veins also are exemplified by their high K₂O/Na₂O ratios (0.63–0.84, adakite <0.5). Another element with anomalous high content compared with typical adakites is TiO₂ (2.20–2.91 wt.%) (Figure S5a). In general, CaO and FeO are higher in low-SiO₂ melts (CaO = 5.82–12.14 wt.%, FeO = 5.11–7.18 wt.%) compared to those of high-SiO₂ (CaO = 4.75–5.73 wt.%, FeO = 2.68–3.46 wt.%) (Figure S5b-c). Samples LG-14 and LG-15 are characterized by the occurrence of two types of melts (high- and low-SiO₂), suggesting the infiltration of a heterogeneous metasomatic agent.

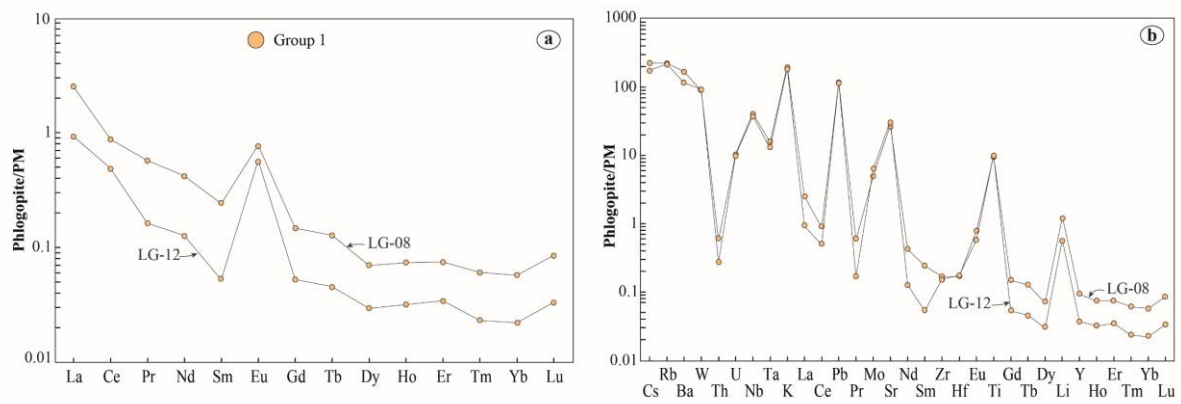


Figure 6. Rare earth element (a) and multielement (b) diagrams of phlogopites from Los Gemelos. The concentrations are normalized to the primitive mantle (PM, Sun & McDonough, 1989).

Associated with these alkaline veins, secondary olivine, orthopyroxene, and clinopyroxene are formed (type 2 minerals). Type 2 olivine has average compositions that indicate a decrease in MgO (45.56–47.00 wt.%, Mg# = 0.86–0.88) and NiO (0.12–0.13 wt.%) along with an increase in FeO (11.60–13.18 wt.%) and CaO (0.23–0.25 wt.%) contents in comparison to primary olivine (MgO = 48.59–50.89 wt.%, Mg# = 0.90–0.92, NiO = 0.37–0.43 wt.%, FeO = 8.27–9.27 wt.%, CaO = 0.02–0.04 wt.%) (Table S2). Type 2 orthopyroxenes do not show significant compositional variation compared to their coexisting type 1 orthopyroxene, although the Mg# tends to be higher in Type 1 (Figure S3). In general, type 2 clinopyroxenes have higher values of TiO₂ (0.06–0.29 wt.%), MgO (18.02–19.22 wt.%), and FeO (2.21–3.15 wt.%), as well as lower SiO₂ (52.00–53.32 wt.%) and CaO (21.04–22.00 wt.%) with respect to the primary phases (TiO₂ = 0.04–0.16, MgO = 15.52–17.36, FeO = 1.80–2.26, SiO₂ = 53.50–54.88, and CaO = 21.44–23.32 wt.%) (Figure 4 and Table S2).

3.5.3. *Pressure and temperature estimates*

Temperature and pressure estimates were obtained assuming 15 kbar (spinel-facies) through several geothermobarometers from the PTEXL Excel spreadsheet (created by T. Köhler in the 1990s and modified and updated by T. Stachel in 2022 at the University of Alberta, Canada: <https://doi.org/10.5683/SP3/IMYNCL>) (Table S9).

Here we employed accurate and robust geothermometers based on two-pyroxenes (Brey & Köhler, 1990; Taylor, 1998), Ca in orthopyroxene (Brey & Köhler, 1990; Nimis & Grütter, 2010), and single-clinopyroxene (Nimis & Taylor, 2000). Based on these results, the temperature average of each sample will be considered hereafter. Reliable pressure estimates were obtained considering the two-pyroxenes model proposed by Brey & Köhler (1990). Therefore, the interval of temperatures for the SCLM beneath Los Gemelos volcano is between 723 and 966 °C, whereas pressure estimates indicate a mantle section between 52 and 74 km (15.6–22.4 kbar) assuming that 10 kbar is equivalent to 33 km (Figure S6). In general, peridotites from G1 equilibrated deeper (21.6 kbar) than those from G2 that represent the intermediate mantle section (19.0–20.3 kbar), whereas most samples from G3 represent the shallower fragments of the SCLM (15.6–18.7 kbar) (Figure S6). Samples LG-04 and LG-16 are stable at higher temperatures (950–966 °C) and pressures (20.1–22.4 kbar) compared to the others from the same compositional group (G3). For comparison pressure and temperature estimates for mantle xenoliths metasomatized by adakite melt metasomatism were plotted in Figure S6. These metasomatized samples are from Cerro del Fraile (Kilian & Stern, 2002) and Paso de Indios region (Cerro Chenque, Cerro Matilde, and Cerro León) (Bertotto *et al.*, 2021) (Figure S6).

3.5.4. *Whole-rock geochemistry*

Whole-rock major and trace element contents were determined for the largest samples (n = 9, Table S4). These results do not allow the individualization of the same compositional groups defined by clinopyroxene mineral chemistry. In terms of major elements, Los Gemelos peridotites contain high-Mg# (0.91–0.92), as well as variable and low CaO (0.74–2.07 wt.%) and Al₂O₃ (0.83–1.80 wt.%) contents (Table S4). The negative correlations observed between MgO vs. CaO, Al₂O₃, Na₂O, and TiO₂ are typical of melt extraction (Figure S7).

In general, all analyzed samples display selective LREE enrichment (e.g., $\text{La}/\text{Sm}_\text{N} = 1.34\text{--}6.31$), and U-shaped REE pattern with variable depletion of MREE relative to LREE and HREE ($\text{La}/\text{Sm}_\text{N} > 1$ and $\text{Gd}/\text{Yb}_\text{N} < 1$; Figure 7a, Table S4). Sample LG-07 has the highest HREE content ($\text{Lu}_\text{N} = 0.41$) and shows a continuous decreasing from Lu to Nd ($\text{Nd}/\text{Lu}_\text{N} = 0.13$) and an increasing from Pr to La ($\text{La}/\text{Pr}_\text{N} = 1.89$). Sample LG-01 also has a U-shaped REE pattern with one of the lowest HREE and the highest LREE contents, with a progressive enrichment from Dy to La ($\text{La}/\text{Dy}_\text{N} = 18.17$). These REE patterns suggest that these samples have undergone variable degrees of partial melting and depletion of the most incompatible elements, followed by selective enrichment of LREE, and MREE, in a few cases.

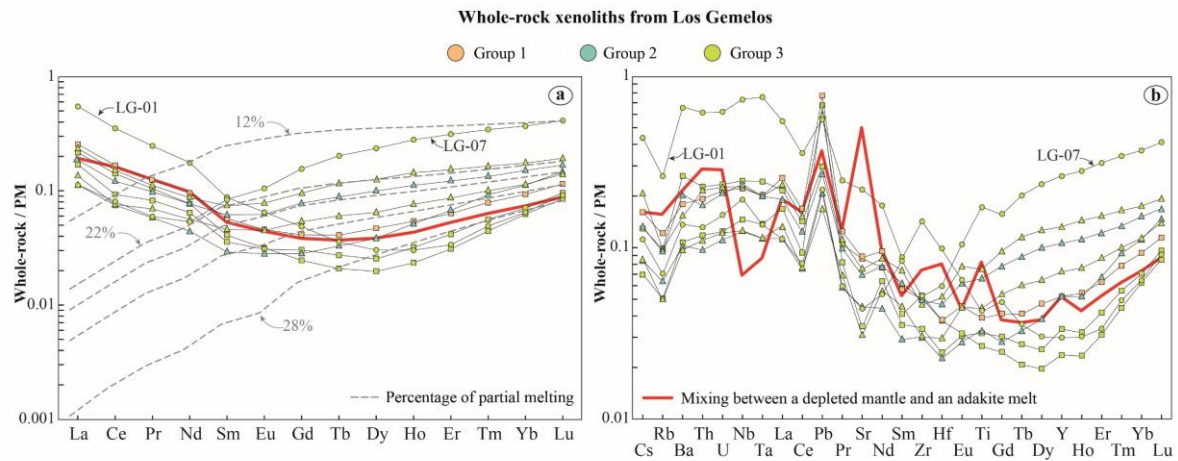


Figure 7. Rare earth element (a) and multielement (b) diagrams of whole-rock samples from Los Gemelos. The concentrations are normalized to the primitive mantle (PM, [Sun and McDonough, 1989](#)). The partial melting estimates (dashed lines) are indicated in 12 %, as well as from 22 to 28 % with the increment of 2 % in (a). Details about the calculations are in the main text. Samples LG01 and LG-07 are highlighted due to their distinct compositional patterns. For comparison, the trace and rare earth elements patterns of a mixing model between a depleted mantle (99.5 %) and an adakite melt (0.5 %; [Corgne et al., 2018](#)) were used to reproduce the compositional patterns of Los Gemelos peridotites (see section 5.1 for details).

Regarding the PM-normalized trace element diagram (Figure 7b), most samples have positive Pb anomalies and smaller positive ones in Cs, Ba, La, and Eu. On the other hand, negative anomalies are observed in Rb, Sr, and Hf (Figure 7b). Sample LG-07 has positive anomalies of Cs, Ba, Nb, Pb, and Ti coupled with negative anomalies of Rb, Sr, Zr, and Hf. Finally, sample LG-01 shows positive anomalies of Cs, Ba, Nb, Ta, Pb, Zr and Hf, coupled with a negative anomaly of Rb (Figure 7b).

3.5.5. Sr-Nd isotopic ratios

Whole-rock Sr-Nd isotopic ratios of Los Gemelos mantle xenoliths, with the exception of samples LG-02 and LG-06, were determined from the same aliquot (rock powder) used for the whole-rock geochemistry (Table S5). The analyzed samples demonstrate moderate $^{87}\text{Sr}/^{86}\text{Sr}$ (0.703882–0.704608) and moderate to low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.512671–0.512836, except for the sample LG-07 with 0.513019) (Figure 8). Thus, in the Sr-Nd diagram (Figure 8), Los Gemelos xenoliths plot between the depleted mantle (DM) and the enriched mantle 1 (EM-1), except for the sample LG-07 that is decoupled from the mantle array and suggest some affinity with the enriched mantle 2 (EM-2) (Hart *et al.*, 1992).

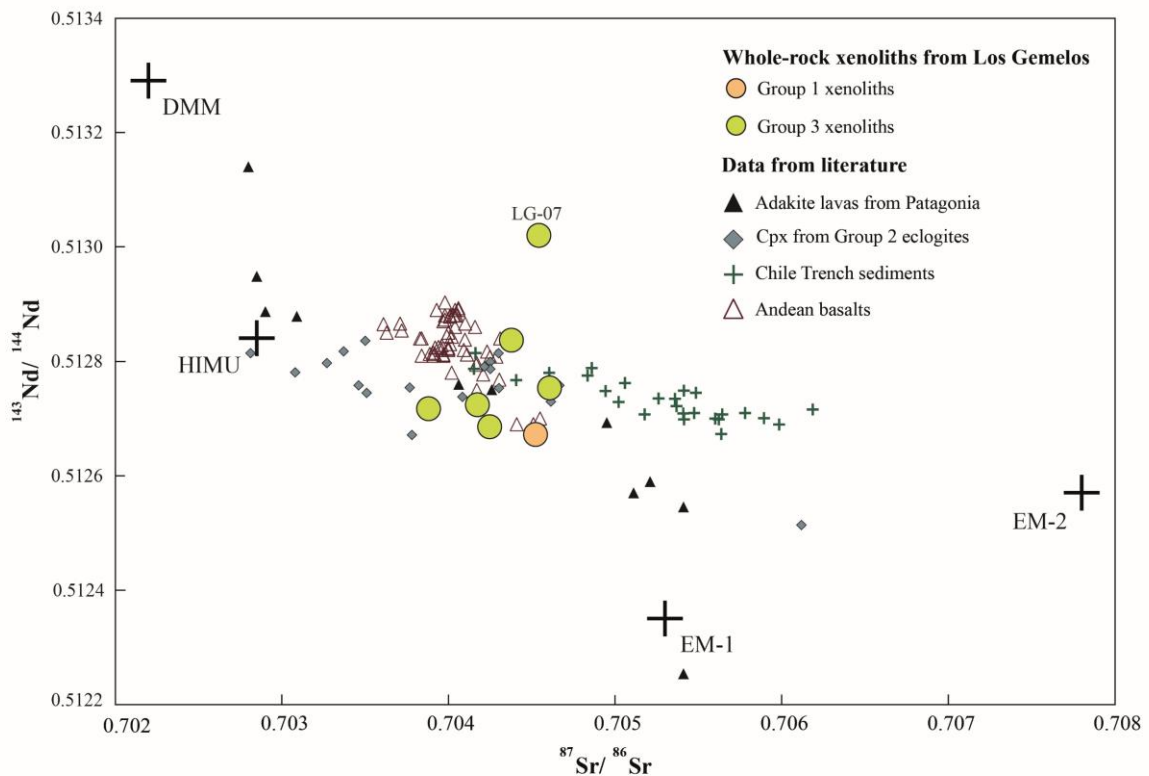


Figure 8. Whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$ diagram of the Los Gemelos peridotites. For comparison, adakite lavas from Cerro Pampa (Mahlburg Kay *et al.*, 1993) and the Austral Volcanic Zone (AVZ, Stern & Kilian, 1996) were plotted. Also, clinopyroxenes from Group 2 eclogites (Viljoen *et al.*, 2005), Andean basalts (Hickey *et al.*, 1986; Hickey-Vargas *et al.*, 1989; D'Orazio *et al.*, 2003), and Chile Trench sediments (Lucassen *et al.*, 2010; Jacques *et al.*, 2013) were used for comparison. Depleted Mantle (DM), Enriched Mantle 1 (EM-1), Enriched Mantle 2 (EM-2), and HIMU mantle reservoirs are from Hart *et al.* (1992). Sample LG-07 is highlighted due to its anomalous result compared to the other samples. In different extents, our samples display compositional affinity with adakite lavas, eclogites, sediments, and continental arc basalts.

3.5.6. *Re-Os and highly siderophile elements*

Considering the limited amount of rock powder, only six whole-rock peridotites from Los Gemelos volcano were analyzed for Re-Os isotopes and HSE contents (Table S6). Initial $^{187}\text{Os}/^{188}\text{Os}$ ratios were calculated considering the eruption age of the host basalt (0.45 ± 0.03 Ma; Table S1), with negligible variations (Table S6). The γ_{Os} (-3.5 to -6.4, where $\gamma_{\text{Os}} = ((^{187}\text{Os}/^{188}\text{Os})_{\text{sample}}/(^{187}\text{Os}/^{188}\text{Os})_{\text{PUM}} - 1) \times 100$; Walker *et al.*, 1989) calculated based on the initial Os ratios, are compatible with Re-depleted residues. The Re-depletion model ages (T_{RD}) of five samples range from 1.1–0.9 Ga (Grenville-age) while sample LG-07 yields a younger age of 0.6 Ga. Also, most samples have mantle model ages (T_{MA}) only slightly older than their T_{RD} model ages, reflective of their low Re/Os ratios. In contrast, samples LG-01 and LG-07 have T_{MA} significantly different than their T_{RD} , suggesting recent addition of Re.

Overall, well-defined positive correlations between $^{187}\text{Os}/^{188}\text{Os}$ vs. whole-rock Al_2O_3 (wt.%) and Yb_N are observed (Figure S8). The $^{187}\text{Os}/^{188}\text{Os}$ vs. Al_2O_3 correlation extrapolates to 0.11819 at $\text{Al}_2\text{O}_3 = 0$, providing a T_{RD} model age of 1.55 Ga, slightly older than the T_{RD} of the individual samples. In contrast, $^{187}\text{Os}/^{188}\text{Os}$ vs. Os (ppb), Ol Mg#, and Sp Cr# define negative correlations (Figure S8). The HSE (PGEs + Re) concentrations (in ppb) of Los Gemelos mantle xenoliths vary widely (Os = 0.024–5.854, Ir = 0.139–1.319, Ru = 0.360–1.510, Pt = 2.468–7.707, Pd = 0.018–0.416, and Re = 0.001–0.017) (Figure 9 and Table S6). Except for the Os_N (1.50) and Pt_N (1.01) contents of LG-06 and LG-01, respectively, all samples are depleted compared to the PUM-normalized HSE values (Becker *et al.*, 2006) (Figure 9). In addition, most samples have similar HSE patterns, with variable depletion of Pd compared to Os, Ir and Ru suggesting different degrees of partial melting, and positive anomalies of Pt (Figure 9 and Table S6). Sample LG-01 shows anomalously low Os and Ir, but with Re concentrations similar to the other xenoliths leading to its anomalously high Re/Os ratio. Sample LG-06 shows enrichment of Os ($\text{Os}_\text{N}/\text{Ir}_\text{N} = 4.35$) (Figure 9 and Table S6). Samples LG-01, LG-03 and LG-08 have relatively high $\text{Re}_\text{N}/\text{Pd}_\text{N}$ ratios (4.7–10.3), also evidencing the subsequent addition of Re after the partial melting process responsible for the Pd depletion.

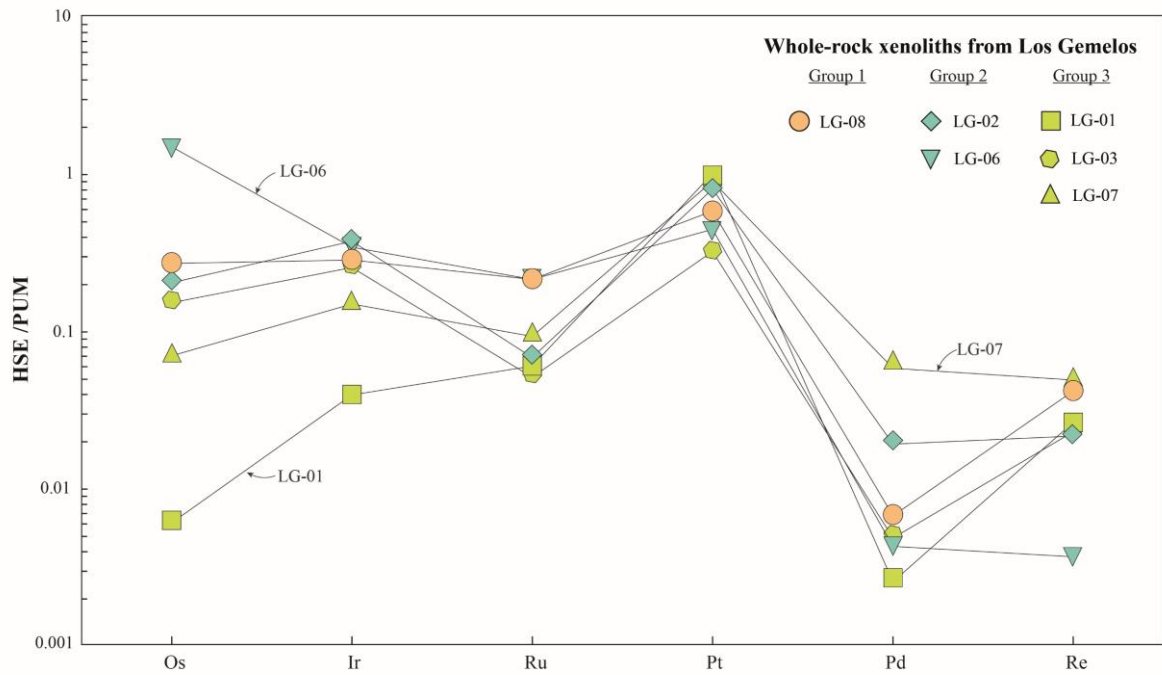


Figure 9. Primitive Upper Mantle (PUM, [Becker *et al.*, 2006](#)) normalized highly siderophile element (HSE) patterns of the Los Gemelos peridotites. Samples LG-01, LG06, and LG-07 are highlighted due to their distinct patterns. In general, our xenoliths present typical partial melting pattern characterized by depletion of Pd and Re coupled with relative enrichment of Os Ir, and Ru. Note that Pt is enriched in all samples.

3.6. Discussion

3.6.1. Partial melting of Los Gemelos mantle xenoliths

Clinopyroxene-poor spinel peridotites from Los Gemelos volcano show petrographic (e.g., presence of alkaline veins and phlogopite) and chemical evidence of a complex mantle evolution involving melt extraction (depletion) followed by enrichment (metasomatism). During the partial melting event, residual peridotites are expected to range from fertile lherzolites to highly refractory clinopyroxene-poor peridotites characterized by low whole-rock Al_2O_3 and CaO contents, high forsterite content in olivine, and Cr-rich spinel. Typical negative correlations between MgO vs. selected basaltic elements (e.g., CaO, Al_2O_3 , Na_2O , and TiO_2), which is expected in the solid residue after partial melting, are observed in whole-rock geochemistry (Figure S7). Most samples also show relative depletion of the more incompatible Re and Pd compared to compatible Ir and Os, consistent with them being residues of melt extraction. The positive $^{187}\text{Os}/^{188}\text{Os}$ vs. Al_2O_3 correlation indicates that this melt extraction occurred well before the event that brought the xenoliths to the surface with T_{RD} ages in the range of a billion years. Still regarding the whole-rock geochemistry, MREE/HREE

ratios support the partial melting process (i.e., $\text{Sm/Yb}_N = 0.23\text{--}0.65$; except for LG-01 = 1.35) (Table S4). Some aspects of mineral chemistry provide further evidence of depletion, as exemplified by the high Cr# of spinels (0.20–0.42) and Mg# of the silicate phases (0.90–0.95) (Table S2). Well-defined negative correlations between Mg# vs. Al_2O_3 and TiO_2 (for both pyroxenes), and CaO vs. Al_2O_3 (for clinopyroxene), besides the positive correlations observed of Mg# vs. CaO and Ca/Al (molar) for clinopyroxene (Figure 4 and Figure S3) attest to extraction of partial melts from these peridotites. Moreover, the negative correlation between Cr# and Mg# of Cr-rich spinels (Figure S2) corroborates this assumption.

Partial melting degrees were estimated based on whole-rock and clinopyroxene HREE compositions (i.e., Yb and Lu) because these elements have similar bulk partition coefficients and, consequently, similar geochemical behavior during partial melting. The extent of partial melting (F) was calculated using the non-modal batch melting (Equation 1 for whole-rock) and the non-modal fractional melting (Equation 2 for clinopyroxene) equations of [Johnson *et al.* \(1990\)](#). They are:

$$C^S = ((D - P \times F) \times C) / (1 - F) \times (D + F \times (1 - P)) \quad (\text{Equation 1})$$

$$\frac{C_i^{\text{cpx}}}{C_i^{\text{o.cpx}}} = \left[1 - \frac{PF}{D_i^{\text{o}}} \right]^{(\frac{1}{P}-1)} \quad (\text{Equation 2})$$

where C^S is the concentration of the element in the solid, D is the distribution coefficient of the element between solid and liquid, F is the fraction of the liquid (melt), and P is the distribution coefficient of the element for the assemblage in equilibrium with the liquid ([Shaw, 2006](#)). As mantle source starting material, we used the primitive mantle REE values ([Sun & McDonough, 1989](#)) and the spinel peridotite modal composition of [McDonough \(1990\)](#) ($\text{Ol} = 0.56$, $\text{Opx} = 0.22$, $\text{Cpx} = 0.19$ and $\text{Sp} = 0.03$). The melt mode for spinel facies ($\text{Ol} = -0.06$, $\text{Opx} = 0.28$, $\text{Cpx} = 0.67$ and $\text{Sp} = 0.11$) is from [Johnson \(1998\)](#) while the mineral/melt partition coefficients were taken from [Ionov *et al.* \(2002\)](#). For whole-rock samples, this modelling yielded between 22 and 28 % of partial melting (Figure 7a). The xenolith LG-07 shows 12 % of partial melting, likely an underestimation compared to the other samples. More likely, this result probably reflects the influence of a later metasomatic event.

Comparatively, the melting model using clinopyroxene requires between 16 and 17 % of partial melting (Figure 5a), which is lower than whole-rock estimates.

3.6.2. *Timing of partial melting*

According with modal, mineral and whole-rock compositions, Los Gemelos peridotites are defined as depleted residues after high degrees of melt extraction (16–28 %) in the spinel-stability field. In this context, HSE and Re are preferentially retained by sulphides. However, at high degrees of partial melting (>20 %) these minerals are completely consumed in the melt (e.g., [Mundl et al., 2016](#) and references therein). As a consequence, the solid residue will be selectively depleted in Re, Pd, and Pt (the latter to a lesser extent) whereas Os, Ir, and Ru will be preferentially retained in alloys in the residual ([Pearson et al., 2004](#)). [Kepezhinskis & Defant \(2001\)](#) associated the selective mantle enrichment of Pt relative to their adjacent elements (Ru and Pd) to the melt extraction in the back-arc of Kamchatka. In this scenario, refractory Pt-Fe alloys selectively concentrate Pt relative to Ru and Pd and, therefore, increase the Pt/Ru and Pt/Pd ratios in the mantle wedge. In this sense, the high PUM-normalized ratios between Os, Ir, and Ru over Re and Pd (values >1) observed in Los Gemelos xenoliths, as well as the apparent enrichment of Pt compared to Ru and Pd, probably reflect the preferential retention of Pt-Fe alloys after high degrees of partial melting (Figure 9). However, there are two exceptions in the tendency mentioned above, where sample LG-01 is enriched in Re relative to Os ($Re_N/Os_N = 4.35$) and sample LG-06 that is enriched in Os with respect to Pt ($Os_N/Pt_N = 3.35$) (Figure 9).

The Re/Os ratio of LG-01 is higher than PM. This is a typical signature of metasomatism by a Re-rich, Os-poor melt whereas the high Os/Pt of LG-06 is the more common signature of melt depletion as Pt is generally more incompatible than Os during mantle melting (e.g., [Pearson et al., 2004](#)). The much lower Re/Os ratios than PM of all samples except LG-01 and LG-07 are a typical signature of melt-depleted peridotites (e.g., [Walker et al., 1989](#)). The sub-PM $^{187}Os/^{188}Os$ isotopic compositions of all samples indicate that this depletion occurred well before the volcanic event that brought these samples to the surface, with T_{RD} model ages averaging 0.97 ± 0.39 (2SD) Ga for all Los Gemelos samples. Leaving out the data for the two samples with higher Re/Os, the average T_{RD} of the remaining samples is 1.09 ± 0.04 Ga. The average T_{MA} of these four samples is 1.22 ± 0.10 Ga, suggesting

that the initial melt depletion of these samples occurred about 1.2–1.1 Ga and that metasomatic addition of Re was insufficient to greatly disturb their Re-Os systematics. Additionally, the positive correlations between $^{187}\text{Os}/^{188}\text{Os}$ vs. whole-rock Al_2O_3 (wt.%) that extrapolates to 0.11819 at $\text{Al}_2\text{O}_3 = 0$, suggests that the partial melting event related to the formation and stabilization of the SCLM beneath Los Gemelos occurred at the beginning of Mesoproterozoic (1.55 Ga).

Sample LG-07 has a Re/Os only slightly lower than the PM value, which might indicate that it was less depleted by melt extraction, though this is not supported by its major and trace element characteristics. The Re/Os ratio of sample LG-01 is much higher than PM, which is inconsistent with its sub-PM $^{187}\text{Os}/^{188}\text{Os}$, and most likely reflects the addition of Re by recent metasomatism. Rough bounds can be placed on when this metasomatism might have occurred by noting that, given its Re/Os ratio, the Os isotopic composition would increase from the lowest measured for any sample from Los Gemelos ($^{187}\text{Os}/^{188}\text{Os} = 0.12134$) to the measured isotopic composition of LG-01 in 60 Ma, which likely provides an upper limit to the time of metasomatism.

3.6.3. *Characteristics and origin of metasomatism*

In this section we will address the metasomatic vestiges recorded by Los Gemelos mantle xenoliths, which are characterized by crystallization of phlogopite, enrichment of clinopyroxene and whole-rock, and occurrence of veins with alkaline composition.

In Patagonia, phlogopite is found in mantle xenoliths from the Gobernador Gregores (Gorring & Kay, 2000; Laurora *et al.*, 2001; Rivalenti *et al.*, 2004; Bjerg *et al.*, 2005) and Pali-Aike Volcanic Field (PAVF, Stern *et al.*, 1999; Jalowitzki *et al.*, 2016). Note that these occurrences experienced the influence of the northward migration of the Chile Ridge from southern South America until its present position at the Taitao Peninsula ($46^\circ 12'S$) since ~16 Ma (Cande & Leslie, 1986; Breitsprecher & Thorkelson, 2009). Thus, the existence of phlogopite as well-crystallized euhedral minerals, as well as mineral inclusions or in the reaction rims around spinels (Figure 2c-e and Figure 3c-d), represents a new occurrence of this hydrated K_2O -rich mineral that attests to effective transfer of volatiles possibly released from the subducted oceanic crust to the overlying mantle wedge. Compositionally, our phlogopites slightly differ from those previously reported for Gobernador Gregores (averages of $\text{MgO} =$

21.44, FeO = 4.69, and Mg# = 0.89; [Gorring & Kay, 2000](#); [Laurora *et al.*, 2001](#); [Bjerg *et al.*, 2005](#)) due to their higher MgO (average of 22.46) and lower FeO (average of 3.32), which imply higher Mg# (average of 0.92). Compared to the other peridotites from Los Gemelos, harzburgites containing phlogopite (G1: LG-08 and LG-12) show distinctive clinopyroxene REE patterns (LREE>HREE) and strong depletions of Nb-Ta, Zr-Hf, and Ti coupled with pronounced enrichment of W, U, Pb, Sr, and Eu (Figure 5).

Aiming to estimate the possible composition of the melt (metasomatic agent) in equilibrium with these clinopyroxenes, we applied the model proposed by [Ionov *et al.* \(2002\)](#), as well as their clinopyroxene/melt partition coefficients. Based on the trace element patterns, the interaction between these clinopyroxenes and the host basalt is discarded (Figure S9a-b). Nevertheless, they share an excellent compositional affinity with natural glass veins and calculated primary melt from Cerro del Fraile ([Kilian & Stern, 2002](#)). They also display strong similarity with the adakite lavas from Cerro Pampa ([Corgne *et al.*, 2018](#)), Southern Chile Trench sediments (i.e., [Kilian & Behrmann, 2003](#)), and with calc-alkaline lavas from Maca and Cay stratovolcanoes ([D'Orazio *et al.*, 2003](#)). Note that the negative anomalies of Nb-Ta and Zr-Hf observed in the potential melt in equilibrium with Los Gemelos clinopyroxenes (Figure S9b) suggest the existence of rutile as a residual phase in the eclogitized oceanic crust during the partial melting ([Foley *et al.*, 2000](#)). The lack of pronounced positive anomaly of Sr in this equilibrated melt probably reflects the contribution of sediments (Figure S9b). The enrichment of Th-U and La-Ce observed in the equilibrated melt compared to subduction-related components (i.e., adakites, calc-alkaline magmas, and sediments) probably is a result of the addition of these incompatible elements to a strongly depleted peridotite.

Moreover, it is important to observe that although amphibole (pargasite) may be a common hydrated mineral phase observed in adakite melt-mantle interaction (e.g., [Rapp *et al.*, 1999](#); [Gervasoni *et al.*, 2017](#); [Corgne *et al.*, 2018](#)), high pressure experiments and natural studies that suggested an interaction of oceanic crust and sediments with the mantle wedge in subduction zones have shown that hydrous K-rich silicate melt reacting with peridotite can produce refertilized zones of hybridized mantle containing phlogopite ([Sekine & Wyllie, 1982a, b](#); [Wyllie & Sekine, 1982](#); [Sekine & Wyllie, 1983](#); [Prouteau *et al.*, 2001](#); [Martin *et al.*, 2005](#); [Grégoire *et al.*, 2008](#); [Förster *et al.*, 2020](#)). Among these experiments, it was suggested that melts derived from sediments infiltrate the mantle wedge and form phlogopite-bearing

pyroxenites at the expense of peridotites following the reaction melt + olivine = phlogopite + pyroxene (Sekine and Wyllie, 1982a, b; Sekine and Wyllie, 1983; Förster *et al.*, 2020).

Phlogopites from Los Gemelos display pronounced positive Eu/Eu* anomalies (4.13–10.88) (Figure 6a), which also is a peculiar feature of phlogopite from Gobernador Gregores (4.39; Gorrington & Kay, 2000), as well as glass veins (1.60) and calculated slab primary melt (1.56) from Cerro del Fraile (Kilian & Stern, 2002). Although less prominent, adakite lavas from the AVZ mainly contain positive Eu/Eu* anomalies (up to 1.27; Mahlborg Kay *et al.*, 1993; Stern & Kilian, 1996; Ramos *et al.*, 2004). On the other hand, calc-alkaline lavas from Cay and Maca stratovolcanoes does not show positive Eu/Eu* anomalies (0.91; D’Orazio *et al.*, 2003). Positive Eu anomalies would be attenuated by the interaction of the slab melt with overlying sediments, producing the enrichment of Eu as a result of replacement of Ca and Sr. This hypothesis was proposed by Stern & Kilian (1996), who invoked the involvement of subducted oceanic crust, sediments, a continental crust component, and the mantle wedge to justify the source of adakite magmas from the AVZ. This model is consistent with our observations, which strongly suggest that phlogopite-bearing harzburgites from Los Gemelos reflects the hybridization of an adakite or calc-alkaline melt, depleted peridotite, and Chile Trench sediment.

In contrast, the estimated melts in equilibrium with clinopyroxenes of G2 and G3 are similar, characterized by depletion from HREE to MREE (Figure S9c). Their compositional patterns differ due to the selective enrichment from Lu to Eu (G2) and from Lu to Pr (G3), in which the former group is more affected by the metasomatism than the latter (Figure S9c). Although we did not find a likely candidate capable of reproduce the trace element patterns of these equilibrated melts, the selective enriched in incompatible elements (i.e., U, Pb, Sr, and Zr) suggests the interaction of a depleted melt (e.g., MORB) with subduction-related components. Based on the information above, we attribute the crystallization of phlogopite, as well as the distinctive clinopyroxene trace element pattern observed in hydrated harzburgites to their deeper origin (71 km) compared to the other xenoliths (52–67 km) (Figure S6). In this way, the intensity of metasomatism decrease toward the shallower SCLM. Notably, a similar scenario was described by Bertotto *et al.*, (2021) to explain the variable extent of metasomatism imposed to mantle xenoliths from Paso de Indios region through the percolation of an adakite melt that promoted melt-rock reactions. Coincidentally, Paso de Indios is located

~70 km northwest from Los Gemelos and both are within in the inferred collision zone between the Deseado and North Patagonian Massif.

In order to achieve a proper discussion concerning the imprint of the metasomatic agents described above on the chemical composition of the Los Gemelos xenoliths, the mineral chemistry results of ortho-(Figure S3) and clinopyroxenes (Figure 4) were compared with typical non-metasomatized (depleted) (Ntaflos *et al.*, 2007; Bertotto *et al.*, 2013, 2022) and adakite metasomatized (Faccini *et al.*, 2013; Bertotto *et al.*, 2021) pyroxenes from other peridotite xenolith localities in Patagonia. Depleted samples are from Agua Poca, Huanul, and Tres Lagos while the metasomatized are from Cerro del Fraile and Paso de Indios region (Cerro León, Cerro Matilde, and Cerro Chenque). We also used for the major and trace elements plots (when these studies contain the selected elements) the pyroxenes metasomatized by adakite melts from an experimental study (Corgne *et al.*, 2018), natural pyroxenes from the Avacha arc (Arai *et al.*, 2003; Ishimaru *et al.*, 2007) and Luzon arc (Arai *et al.*, 2004), Group 2 eclogites (Viljoen *et al.*, 2005), and reaction products (type 2 secondary minerals) from alkaline veins of Los Gemelos mantle xenoliths.

Depleted pyroxenes should be characterized by high Mg# (usually >90) coupled with low concentrations of basaltic elements (except for CaO that is enriched in clinopyroxene). However, the metasomatism caused by the percolation of adakite melts produces similar effect (Figure 4 and Figure S3). For this reason, although pyroxenes from G2 and G3 plot within the fields of samples affected by adakite metasomatism (Figure 4 and Figure S3), we prefer to be careful with the interpretation of these results. This assumption is justified by the absence of phlogopite in these samples (G2 and G3), as well as by the significant difference of their clinopyroxenes REE and extended trace element patterns compared to those from G1 (Figure 5). Hence, afterwards we will use the major elements of pyroxenes as a reliable indicator of adakite (or calc-alkaline) metasomatism only to phlogopite-bearing harzburgites (samples LG-08 and LG-12).

In selected major element diagrams of clinopyroxenes (Mg# vs. Al_2O_3 , TiO_2 , CaO, and Ca/Al; as well as Al_2O_3 vs. CaO) and orthopyroxenes (Mg# vs. Al_2O_3 , TiO_2 , and CaO) phlogopite-bearing harzburgites show similar composition with xenoliths metasomatized by adakite melts from Patagonia, Avacha, and Luzon (Figure 4 and Figure S3). Moreover, these pyroxenes (type 1) overlap the same compositional fields of type 2 from glass veins. In general, these samples also plot within the experimental results of pyroxenes characterized by

enrichment promoted by the interaction with adakite melt, although these experiments have slightly higher Al_2O_3 and TiO_2 coupled with lower CaO contents. In these diagrams, clinopyroxenes from hydrated harzburgites from Los Gemelos also display compositional similarity with Group 2 eclogites, giving additional evidence of their formation through the interaction with a slab-derived melt from a hydrous eclogitic protolith (adakite) clearly promoting low Al_2O_3 and TiO_2 accompanied by high CaO and Ca/Al at a given high-Mg# (Figure 4).

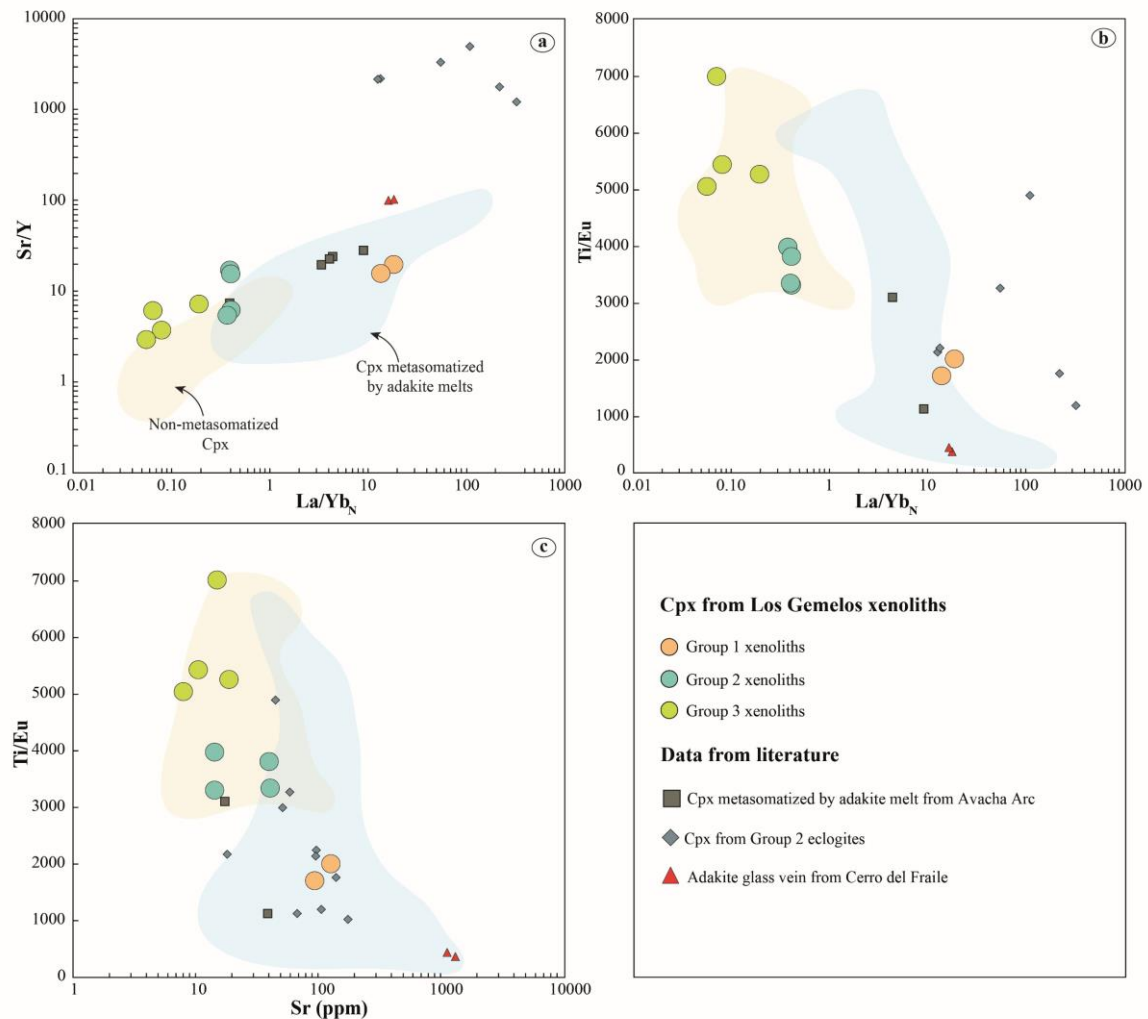


Figure 10. La/Yb_N vs. Sr/Y (a) and Ti/Eu (b), and Sr (ppm) vs. Ti/Eu (c) diagrams of clinopyroxenes (Cpx) from Los Gemelos peridotites. Notably, only Group 1 samples (those containing phlogopite) record clear compositional affinity with clinopyroxenes metasomatized by adakite melt from Avacha (Ishimaru *et al.*, 2007) and from Group 2 eclogites (Viljoen *et al.*, 2005), as well as with glass veins with adakite composition from Cerro del Fraile (Kilian & Stern, 2002). Group 2 clinopyroxenes from Los Gemelos show less evident correlation, whereas those from Group 3 lack compositional affinity with metasomatic process. This observation defines decoupling between major and trace elements (see Figure 3 for comparison and the main text for discussion). The fields defined by non-metasomatized (light yellow) and metasomatized (light blue) peridotites from Patagonia are the same as in Figure 3. The La/Yb ratio is normalized ($_N$) to the primitive mantle (Sun & McDonough, 1989).

Regarding trace and rare earth elements of clinopyroxenes, Figure 10 correlates La/Yb_N vs. Sr/Y (a) and Ti/Eu (b), as well as Sr (ppm) vs. Ti/Eu (c). The selection of these elements is justified due to the preferential enrichment of La, Sr and Eu coupled with depletion of Yb, Y, and Ti during the metasomatism promoted by the percolation of eclogite-derived adakite melt into the mantle wedge. In these diagrams, the strong affinity of G1 clinopyroxenes from Los Gemelos (with coexisting phlogopite) with adakite melts is confirmed. They plot within metasomatized clinopyroxenes from Patagonia (Faccini *et al.*, 2013; Bertotto *et al.*, 2021) and the Avacha arc (Ishimaru *et al.*, 2007), sometimes overlapping or towards the compositions of glass veins from Cerro del Fraile (Kilian & Stern, 2002) and Group 2 eclogites (Viljoen *et al.*, 2005) (Figure 10). Due to their less pronounced enriched character regarding incompatible trace elements, clinopyroxenes from G2 and G3 mainly resemble the field of depleted peridotites from Patagonia (Ntaflos *et al.*, 2007; Bertotto *et al.*, 2013, 2022) (Figure 10). This observation suggests that these two compositional groups were less affected by the adakite (or calc-alkaline) metasomatism.

Regarding the whole-rock geochemistry, usually distinctive elemental characteristics are promoted by reaction between adakite melts and peridotites, such as the enrichment of key elements (i.e., Pb, Sr, Zr, Hf, and LREE/HREE) and depletion of Nb-Ta (e.g., Defant & Drummond, 1990; Stern & Kilian, 1996; Kilian & Stern, 2002; Martin *et al.*, 2005; Fallon *et al.*, 2008; Corgne *et al.*, 2018). In general, Los Gemelos peridotites with U-shaped REE patterns lack the typical positive anomalies of Sr, Zr, and Hf, as well as negative anomalies of Nb-Ta, characteristics of adakites (except for few samples with negative Ta and positive Zr anomalies) (Figure 7b). The simplest explanation for the U-shaped REE pattern of these peridotites is that they were metasomatized by a LREE-enriched alkaline (adakite?) melt after high degrees of partial melting. Therefore, the addition of a low amount of this LREE-enriched melt to a strongly LREE-depleted pattern generates the selective enrichment of LREE abundances, whereas the MREE and HREE will maintain the signature of past melt depletion. Aiming to reproduce the U-shaped REE pattern of Los Gemelos peridotites, we modelled a mixing between a depleted mantle after 28 % of partial melting from a PM composition (see section 5.1) and an adakite melt from Cerro Pampa (Corgne *et al.*, 2018) (in the proportion of 99.5 and 0.5 %, respectively) (Figure 7). The REE pattern of these endmembers match that expected for depleted peridotites metasomatized by small amount adakite melt, as observed in

the whole-rock composition of the Los Gemelos peridotites (Figure 7a). On the other hand, the trace element pattern of this modeling presents the typical compositional of adakite metasomatism (Figure 7b). Hence, the metasomatic overprint in the Los Gemelos peridotites was enough to significantly add LREE to a starting depleted peridotite to change the REE pattern, but not enough metasomatic agent was added to significantly influence the abundance of all these other elements, which would not be as depleted as LREE in a depleted peridotite. Therefore, the lack of positive anomalies of Sr, Zr, and Hf coupled with negative anomalies of Nb-Ta probably reflects an inherited characteristic that was not effectively modified by the adakite melt metasomatism.

Still considering the whole-rock geochemistry, samples LG-01 and LG-7 clearly have distinct patterns (Figure 7), probably implying a complex and different interaction between metasomatic agent and residual peridotite. Although these two samples do not exhibit modal metasomatism (with no phlogopite and/or adakite glass veins), their higher Re/Os suggests that they contain more of the metasomatic component than do the other samples.

Isotopically, Los Gemelos peridotites plot within the Sr-Nd isotopic ratios of adakite lavas from Cerro Pampa ([Mahlburg Kay *et al.*, 1993](#)) and the AVZ ([Stern & Kilian, 1996](#)), resembling the field of clinopyroxene separates from Group 2 eclogites ([Viljoen *et al.*, 2005](#)) (Figure 8). They partially overlap the more enriched Andean basaltic samples ([Hickey *et al.*, 1986](#); [Hickey-Vargas *et al.*, 1989](#); [D’Orazio *et al.*, 2003](#)) and the less enriched samples from the Chile Trench sediments ([Lucassen *et al.*, 2010](#); [Jacques *et al.*, 2013](#)) (Figure 8). Although the Sr and Nd isotopic composition of Los Gemelos xenoliths might represent an intrinsic less depleted character than observed in MORB mantle sources, we assume that they record the addition of a small fraction of slab-derived components. This assumption is corroborated by several chemical evidence that define these samples as highly refractory peridotites and, therefore, their isotopic ratios if unaffected by metasomatism, should record a depleted character consistent with their high degrees of partial melting.

Another expression of metasomatism recorded by mantle xenoliths from Los Gemelos is the occurrence of crosscutting networks of veins apparently filling hydraulic fractures within mineral phases (LG-02, LG-06, LG-08, LG-12, LG-14, LG-15) (Figure 2b, f and Figure 3e-f; Table S2). Compositionally, these veins represent the quenching of strong alkaline melts (7.37–9.59 wt.%) characterized by high-K contents (2.84–4.37 wt.%). They can be classified as high-SiO₂ (>60 %, HSA: SiO₂ >60 wt.%) and low-SiO₂ adakite melts (<51 %, LSA, SiO₂

<60 wt.%) (Martin *et al.*, 2005) (Figure S4 and Figure S5). Due to their higher K₂O and TiO₂, as well as slightly lower Al₂O₃, CaO, and MgO at a given SiO₂ content, high-SiO₂ glass veins from Los Gemelos differ from the adakite lavas from Patagonia (Mahlburg Kay *et al.*, 1993; Stern & Kilian, 1996; Ramos *et al.*, 2004; Corgne *et al.*, 2018), experimental studies (Rapp *et al.*, 1999; Corgne *et al.*, 2018), as well as from adakite glasses veins in metasomatized peridotites from Cerro del Fraile (Kilian & Stern, 2002) (Figure S4 and Figure S5). Regarding Na₂O, our samples overlap the samples mentioned above. This comparison can be extended to adakite glasses of metasomatized peridotites from Avacha and Luzon arcs (Arai *et al.*, 2003, 2004; Ishimaru *et al.*, 2007) (Figure S4 and Figure S5). Contrary, the low-SiO₂ veins from Los Gemelos show strong affinity with low-SiO₂ adakite lavas from the East Coast rifted margin from the United States of America (Meyer & Wijk, 2015) (Figure S4 and Figure S5). Independent of SiO₂ content, the Mg# of veins from Los Gemelos is high (0.56–0.64), which indicates contribution of a peridotitic component.

Experimentally, the presence of CO₂ in eclogite-derived partial melts reduces SiO₂ and Al₂O₃ contents and increases FeO, CaO and CaO/Al₂O₃ (Gerbode & Dasgupta, 2010; Kiseeva *et al.*, 2012). Comparatively, the low-SiO₂ melts from Los Gemelos have higher FeO (5.11–7.18 wt.%), CaO (5.82–12.14 wt.%), and CaO/Al₂O₃ (0.34–0.73) coupled with lower Al₂O₃ (16.56–17.62 wt.%) than high-SiO₂ melts (CaO = 4.75–5.73 wt.%, FeO = 2.68–3.46 wt.%, CaO/Al₂O₃ = 0.29–0.35, Al₂O₃ = 15.78–16.52 wt.%) (Figure S4 and Figure S5, not all values are shown).

The lack of reaction zones in the peridotitic paragenesis associated with percolation of glass veins, as well as the expectation that at mantle conditions they would be re-equilibrated with the surrounding peridotites during long lithospheric residence, we presume that the veins were formed by relatively recent/young processes, close to the time of eruption of the host basalt.

3.6.4. Implications for tectonic Evolution

Mantle xenoliths from Los Gemelos were transported to the surface by its host lava during the Pleistocene (0.45 ± 0.03 Ma, Table S1). They record relatively high degrees of partial melting (16–28 %) that occurred during the Mesoproterozoic (1.6–1.0 Ga). The Grenville-ages of most Los Gemelos peridotites (T_{RD} and T_{MA} = 1.3–1.1 Ga) can be associated

to the formation and stabilization of the Patagonian SCLM that occurred during the Rodinia amalgamation. [Pankhurst *et al.* \(2014\)](#) reported that in this region zircon cores with Grenville-age have Hf model ages between 1.4 and 1.3 Ga and juvenile ϵHf (+7.1 to +8.2), confirming the depleted mantle source contribution to the initial formation of continental crust in this area. The Re-Os isotopic compositions and T_{RD} model ages obtained for Los Gemelos (1.13–0.63 Ga) are similar to previous results obtained for mantle xenoliths from the Deseado Massif (T_{RD} ages = 1.34–0.50 Ga, $n = 9$) and the North Patagonian Massif (T_{RD} ages = 1.30–0.43 Ga, $n = 23$) ([Schilling *et al.*, 2008, 2017](#); [Mundl *et al.*, 2015, 2016](#)). However, older ages also were reported for mantle xenoliths from the Deseado Massif (T_{RD} ages = 2.13–1.52 Ga, $n = 8$). Based on these Paleo- to Mesoproterozoic ages, [Schilling *et al.* \(2017\)](#) have related the formation of the SCLM beneath the Deseado and North Patagonian massifs to the Malvinas/Falkland Islands and to the Namaqua-Natal province (southern South Africa) (e.g., [Thomas *et al.*, 2000](#); [Wareham *et al.*, 1998](#)). Therefore, we conclude that the Los Gemelos peridotites provide additional evidence about the participation of the Patagonia together with other continental blocks in the amalgamation of Rodinia supercontinent, which is contemporaneous to the Grenville orogeny.

After the crustal formation of Patagonia during the Proterozoic, different tectonic events can be considered potentially responsible to explain the modal and cryptic metasomatism shown by Los Gemelos peridotites. Considering that the generation of adakite melts requires high temperature (heat influx), these events must involve the development of slab-windows and asthenospheric upwelling. The subduction of young and hot oceanic crust related to the Chile Ridge under South America started ~16 Ma ago at the southern edge of the continent and has moved to its present location near the Taitao Peninsula (46 °S). This tectonic configuration resulted in the formation of adakitic lavas erupted at six active stratovolcanoes of the Austral Volcanic Zone (AVZ) during the Holocene (Lautaro, Viedma, Aguilera, Reclus, Mt. Burney, and Cook Island; [Stern & Kilian, 1996](#)). Also, Miocene adakite igneous rocks were discovered further east (Cerro Pampa, Puesto Nuevo, and Chaitén: average of 12.50 ± 1.13 Ma; [Ramos *et al.*, 1991, 2004](#); [Mahlburg Kay *et al.*, 1993](#); [Motoki *et al.*, 2003](#)). Notably, Chaitén, Puesto Nuevo, and Cerro Pampa adakites record northwards decreasing eruption ages (14.50 ± 0.29 , 13.12 ± 0.55 , 11.39 ± 0.61 Ma, respectively), consistent with melting of the trailing edge of the Nazca plate as ridge collision progressed to the north ([Ramos *et al.*, 2004](#)). However, adakites were not described in the Southern Volcanic Zone (SVZ) or

even in its back-arc region. Finally, mantle xenoliths from Cerro del Fraile brought to the surface by Pleistocene alkaline basalts (2.44–1.01 Ma, [Fleck *et al.*, 1972](#)) are characterized by adakite metasomatism ([Kilian & Stern, 2002](#); [Faccini *et al.*, 2013](#)). However, considering the timing and evolution of the Chile Ridge subduction, this cannot be invoked as the responsible process for the metasomatism recorded by Los Gemelos mantle xenoliths, which is located around 600 km to the northeast of the Chile Triple Junction. In this tectonic context, the only alternative to justify the interaction of an adakite melt with Los Gemelos peridotites during the Pleistocene is the partial melting of a remnant slice of an ancient (probably from Paleocene) eclogitized oceanic crust immediately before or concomitantly with the ascent of the host magma (proto-basalt metasomatism?).

Another possibility to explain the mantle metasomatism is related to tectonic processes that occurred during the Paleocene to Eocene. Plate tectonic reconstruction (e.g., [Cande & Leslie, 1986](#); [Breitsprecher & Thorkelson, 2009](#)) associated with geochronological and geochemical data for Paleocene-Eocene flood basalts outcropping in the Aysén region (46°S: 60–46 Ma; [Baker *et al.*, 1981](#); [Butler *et al.*, 1991](#); [Demant *et al.*, 1996](#); [Morata *et al.*, 2000](#); [Parada *et al.*, 2001](#); [Jalowitzki *et al.*, 2017](#)) attest to the southward passage of the Farallón-Aluk trench-ridge-trench triple junction at ~60–50 Ma ([Jalowitzki *et al.*, 2017](#)). Similarly, the Paso de Indios region (~43°50'S) is composed mainly by coeval Paleocene-Eocene lavas (~60–50 Ma, [Alric, 1996](#); [Haller *et al.*, 2020](#) and references therein). The estimate of the time of Re-enrichment seen in sample LG-01 also is ~60 Ma. However, there is no evidence for Paleocene-Eocene magmatism in the Miocene Canquel Plateau (23–21 Ma; [Anselmi *et al.*, 2004](#) and references therein). Tectonically, the main-plateau stage of this magmatism has been related to the physical perturbation produced by the arrival of the detached oceanic slab (Farallón) into the hydrated mantle transition zone, which promoted the breakdown of wadsleyite and favored the release of water (wet plumes) that triggered the partial melting of shallower upper mantle ([Honda *et al.*, 2006](#); [Orihashi *et al.*, 2006, 2010](#); [Navarrete *et al.*, 2020](#)). As the Canquel and Somuncura plateaus do not record the magmatic activity during the Paleocene-Eocene related to the slab-window, we consider it likely that at this time, the SCLM beneath Los Gemelos has interacted with adakite melts released from the edge of the relatively hot oceanic plate including sediments.

3.7. Conclusion

Here we report new insights about the complex and long-lived SCLM evolution beneath the Patagonian back-arc by examining spinel-bearing mantle xenoliths from the Los Gemelos volcano, Canquel Plateau. They are mainly harzburgites ($n = 14$) and few clinopyroxene-poor lherzolites ($n = 2$) characterized by high-degrees of melt extraction (16–28 %) recording superimposed enrichment (metasomatism). The depleted character is evidenced by low whole-rock Al_2O_3 and CaO contents (both <1.5 wt.%), high Mg# olivine (>90), and Cr-rich spinel ($\text{Cr\#} >0.20$), together with correlations between Mg# vs. selected basaltic elements (e.g., CaO, Al_2O_3 , Na_2O , and TiO_2) in whole-rock and pyroxenes. In addition, HSE (PGEs + Re) patterns support this assumption with relative depletions in the incompatible elements Re and Pd. Re-Os model ages and the positive correlations between $^{187}\text{Os}/^{188}\text{Os}$ vs. whole-rock Al_2O_3 (wt.%) suggest that the partial melting event related to the formation and stabilization of the SCLM beneath Los Gemelos occurred during the Mesoproterozoic (1.6–1.0 Ga).

The petrographic and chemical characteristics of the SCLM beneath Los Gemelos volcano reflects modal and cryptic metasomatism. The modal metasomatism is evidenced by the crystallization of phlogopite and by the abundant occurrence of well-developed crosscutting networks of glass veins with alkaline composition containing low- and high- SiO_2 . Also, cryptic metasomatism is evidenced by the selective enrichment of LREE (La-Ce) compared to the other REEs in clinopyroxene ($\text{La}/\text{Pr}_N = 1.50\text{--}6.89$), as well as by the U-shaped REE pattern shown by whole-rock compositions ($\text{La}/\text{Sm}_N >1$ and $\text{Gd}/\text{Yb}_N <1$). Selected discrimination diagrams including major (i.e., Mg# vs. basaltic elements) and trace (i.e., La/Yb_N vs. Sr/Y and Ti/Eu ; and Sr vs. Ti/Eu) elements of pyroxenes strongly suggest the role of a silicate melt with adakite composition as a likely candidate for the metasomatic agent responsible for the enrichment (in variable extents) of the SCLM beneath Los Gemelos volcano. Notably, the intensity of metasomatism decreases from the deeper (~ 70 km) towards the shallower (~ 50 km) mantle section. The calculated melts in equilibrium with clinopyroxenes revealed that the G1 (with coexisting phlogopite) were strongly metasomatized by an adakite melt, whereas G2 and G3 were slightly affected by this metasomatic event. Although not conclusive, whole-rock Sr-Nd isotopes suggest the addition of typical subduction-related components. Therefore, it is reasonable to conclude that the crystallization

of phlogopite and cryptic metasomatism recorded by mantle xenoliths is a product of partial melting of subducted hot oceanic crust that triggered adakite melts as metasomatic agents in the mantle wedge. Among the probable tectonic events responsible for the adakite metasomatism identified in Los Gemelos mantle xenoliths, we consider the development of a slab-window during the Paleocene to Eocene as a preferred candidate.

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MATERIAL SUPPLEMENTAR

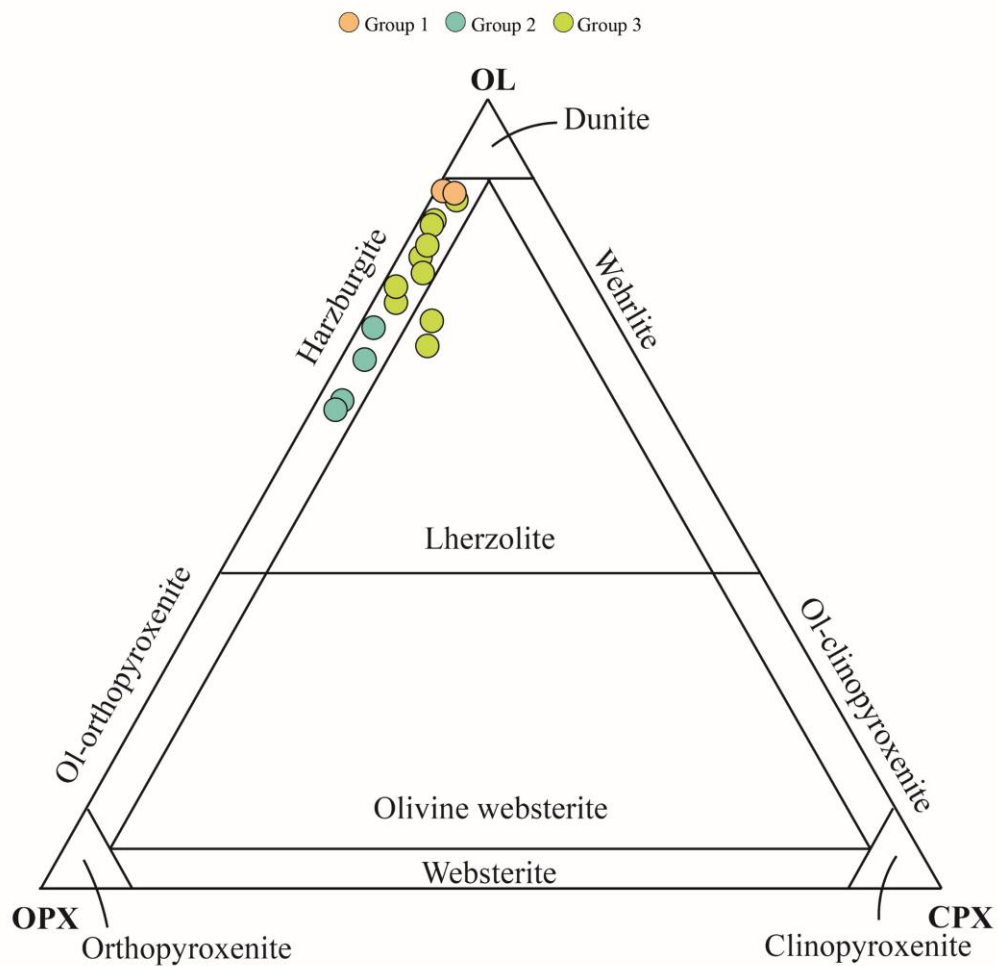


Figure S1. Ternary diagram correlating the percentages of olivine (OL), orthopyroxene (OPX), and clinopyroxene (CPX) of mantle xenoliths from Los Gemelos. Most of the studied samples are classified as harzburgites, while LG-01 and LG-10 are clinopyroxene-poor lherzolites.

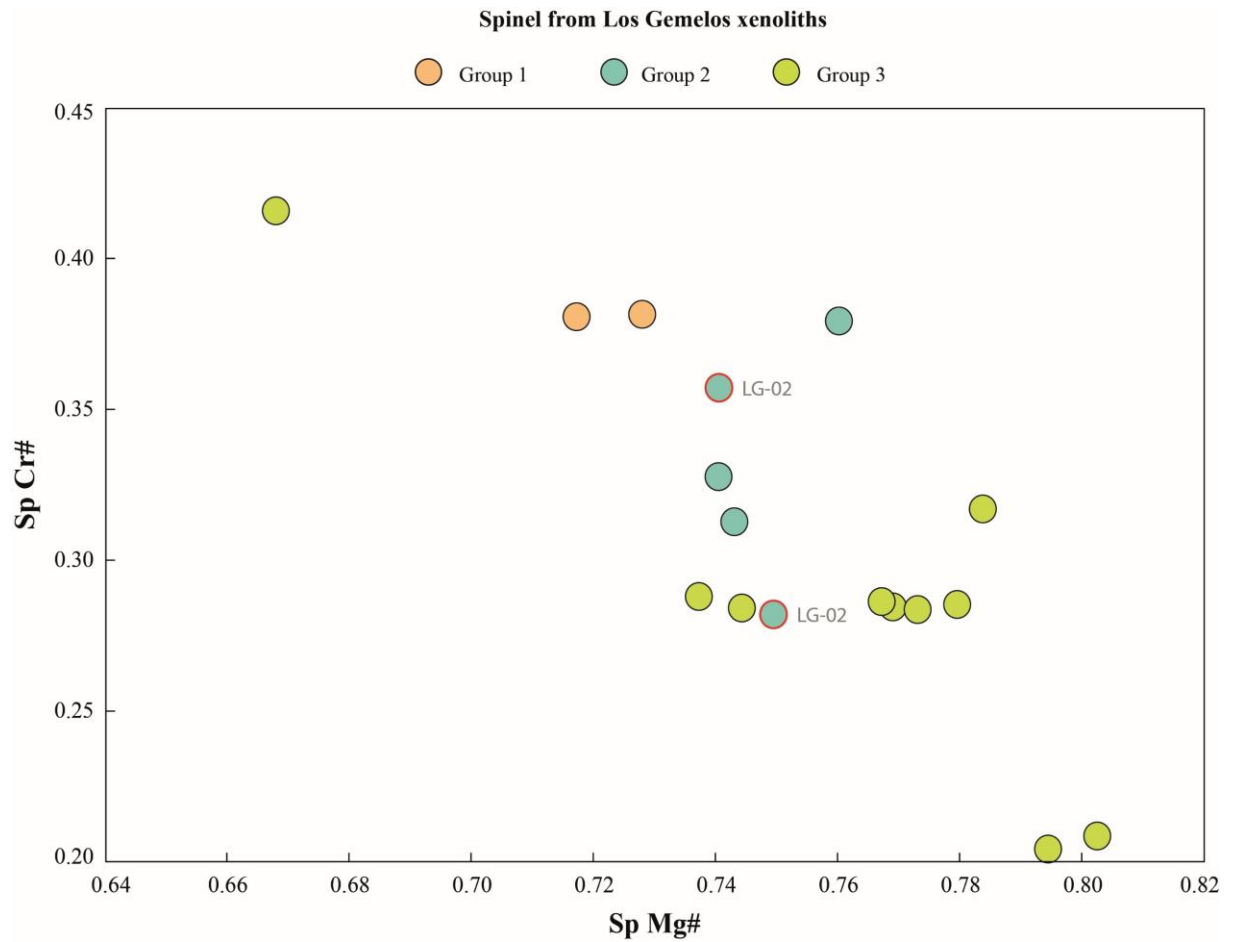


Figure S2. Spinel Mg# vs. Cr# of mantle xenoliths from Los Gemelos showing negative correlation, typical of partial melting. Sample LG-02 is highlighted because it is the only sample showing two compositions of spinel.

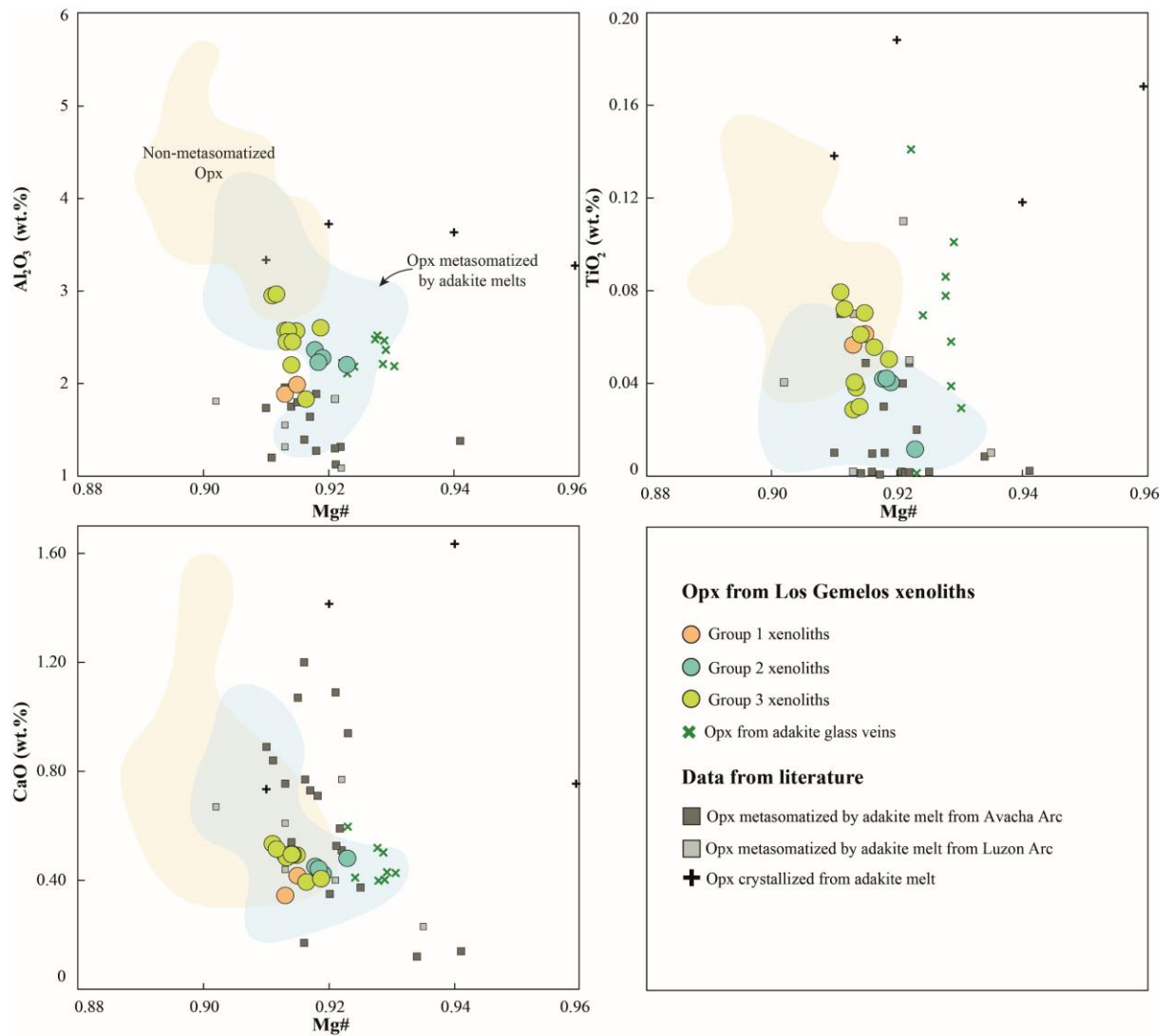


Figure S3. Selected major elements (Al₂O₃, TiO₂, and CaO) vs. Mg# (a-c) of orthopyroxenes (Opx) classified as primary (type 1) and crystallized within adakite glass veins (type 2) from Los Gemelos. Our samples show compositional affinity with orthopyroxenes metasomatized by adakite melts (light blue field) of mantle xenoliths from Cerro de los Frailes and the Paso de Indios region (Faccini et al., 2013; Bertotto et al., 2021). The field defined for non-metasomatized orthopyroxenes from Patagonia (light yellow) includes mantle xenoliths from Agua Poca, Huanul, and Tres Lagos (Ntaflos et al., 2007; Bertotto et al., 2013, 2022). For comparison, we plotted orthopyroxenes metasomatized by adakite melts from the Avacha and Luzon arcs (Arai et al., 2003, 2004; Ishimaru et al., 2007), and experimentally crystallized from adakite melts (Corgne et al., 2018).

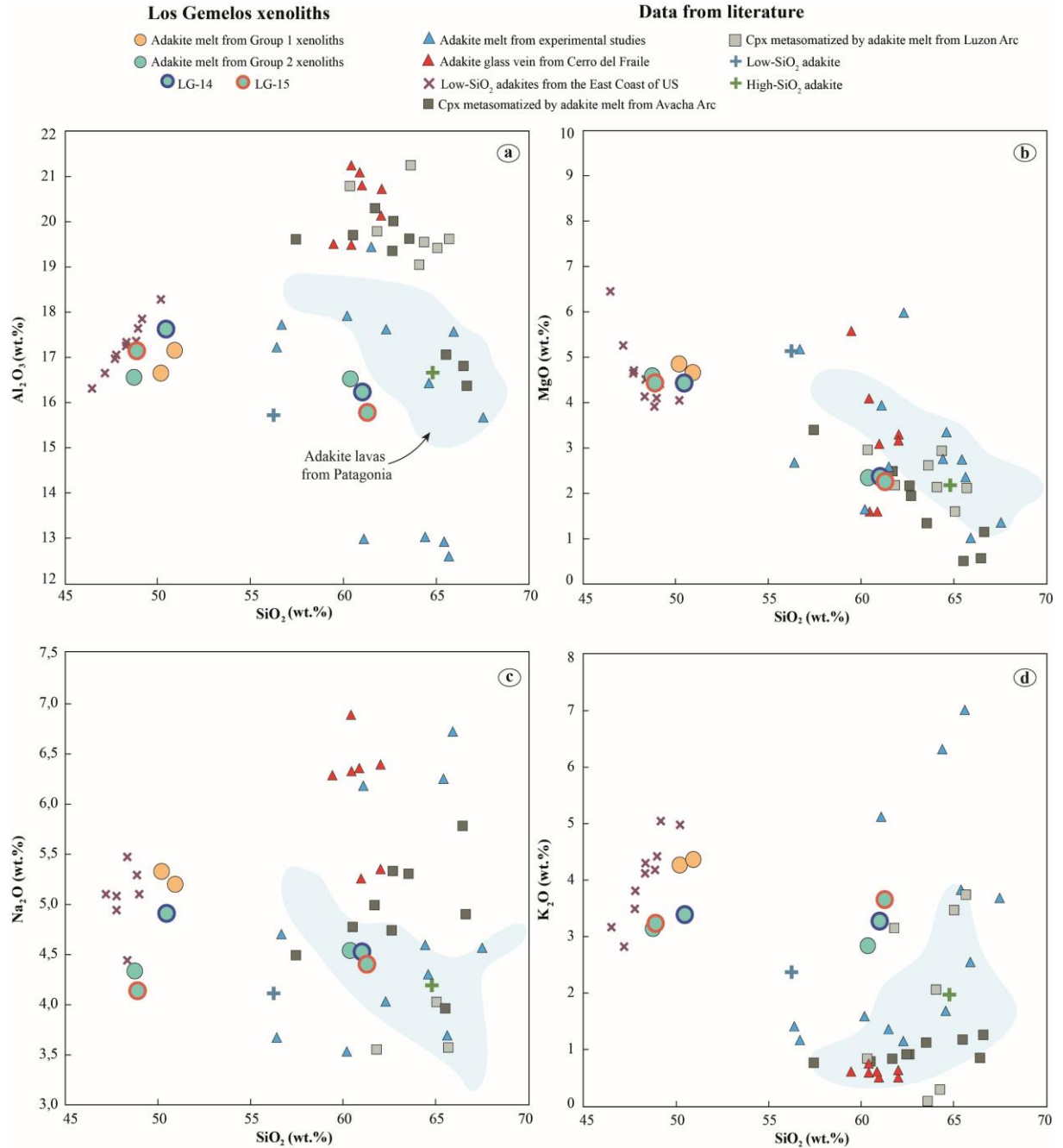


Figure S4. Selected major elements (Al₂O₃, MgO, Na₂O, and K₂O) vs. SiO₂ of glass veins with adakite composition from Los Gemelos (a-d). Our samples show strong compositional variation from low- to high-SiO₂ adakite melts, in which mantle xenoliths from LG-14 and LG-15 present both compositions. For comparison, adakite lavas from Patagonia (Stern and Kilian, 1996; Mahlborg Kay et al., 1993; Ramos et al., 2004; Corgne et al., 2018), adakite melts from experimental (Rapp et al., 1999; Corgne et al., 2018) and natural (Cerro del Fraile, Kilian and Stern, 2002) samples, clinopyroxenes metasomatized by adakite melts from Avacha and Luzon arcs (Arai et al., 2004; Ishimaru et al., 2007), and low-SiO₂ adakites from the East Coast of United States of America (Meyer and Wijk, 2015) were plotted. In addition, the averages of low- and high-SiO₂ adakites (Martin et al., 2005) were plotted.

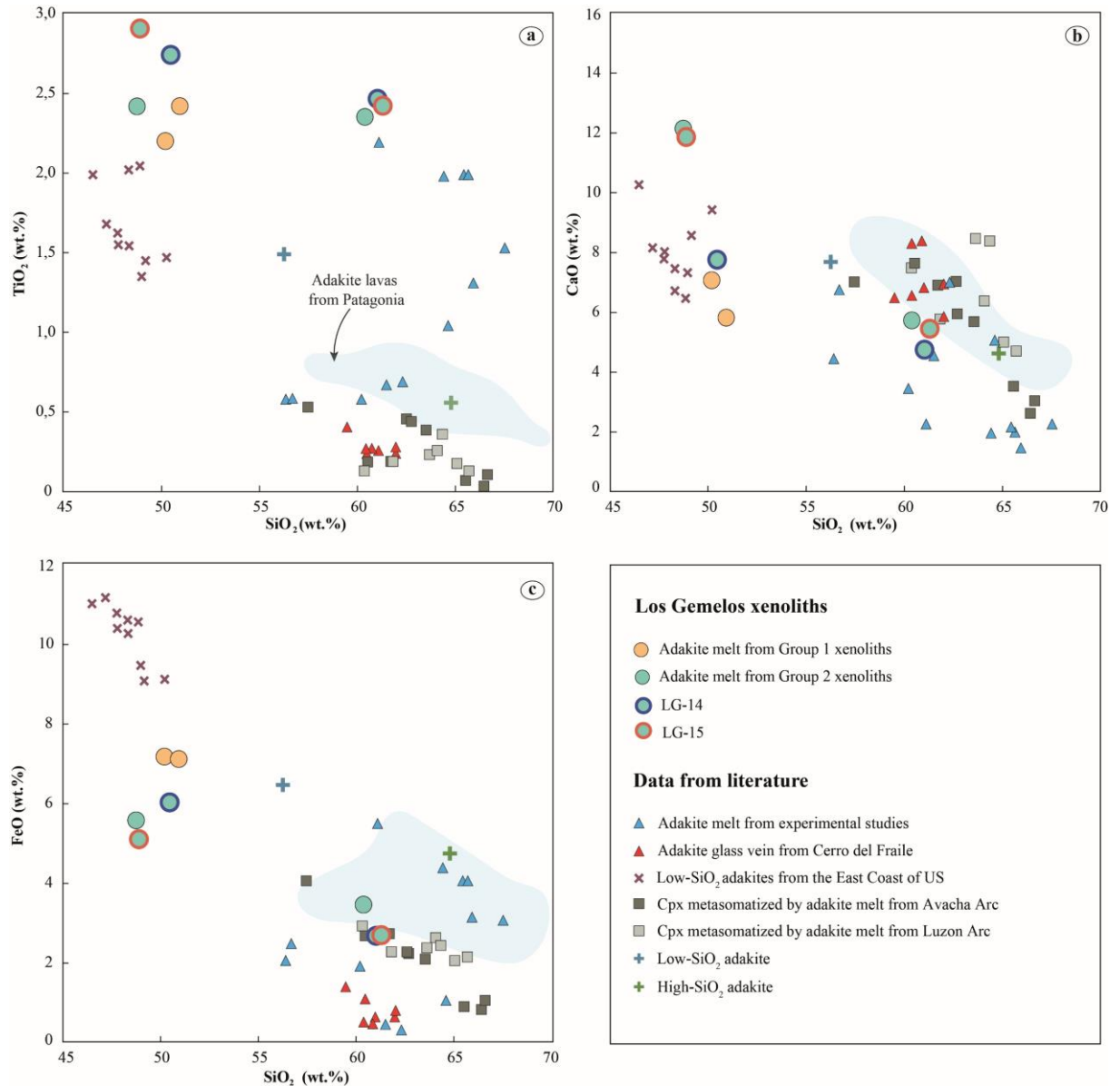


Figure S5. Selected major elements (TiO₂, CaO, and FeO) vs. SiO₂ of glass veins with adakite composition from Los Gemelos (a-c). Our samples show strong compositional variation from low- to high-SiO₂ adakite melts, in which mantle xenoliths from LG-14 and LG-15 present both compositions. For comparison, adakite lavas from Patagonia (Stern and Kilian, 1996; Mahlburg Kay et al., 1993; Ramos et al., 2004; Corgne et al., 2018), adakite melts from experimental (Rapp et al., 1999; Corgne et al., 2018) and natural (Cerro del Fraile, Kilian and Stern, 2002) samples, clinopyroxenes metasomatized by adakite melts from Avacha and Luzon arcs (Arai et al., 2004; Ishimaru et al., 2007), and low-SiO₂ adakites from the East Coast of United States of America (Meyer and Wijk, 2015) were plotted. In addition, the averages of low- and high-SiO₂ adakites (Martin et al., 2005) were plotted.

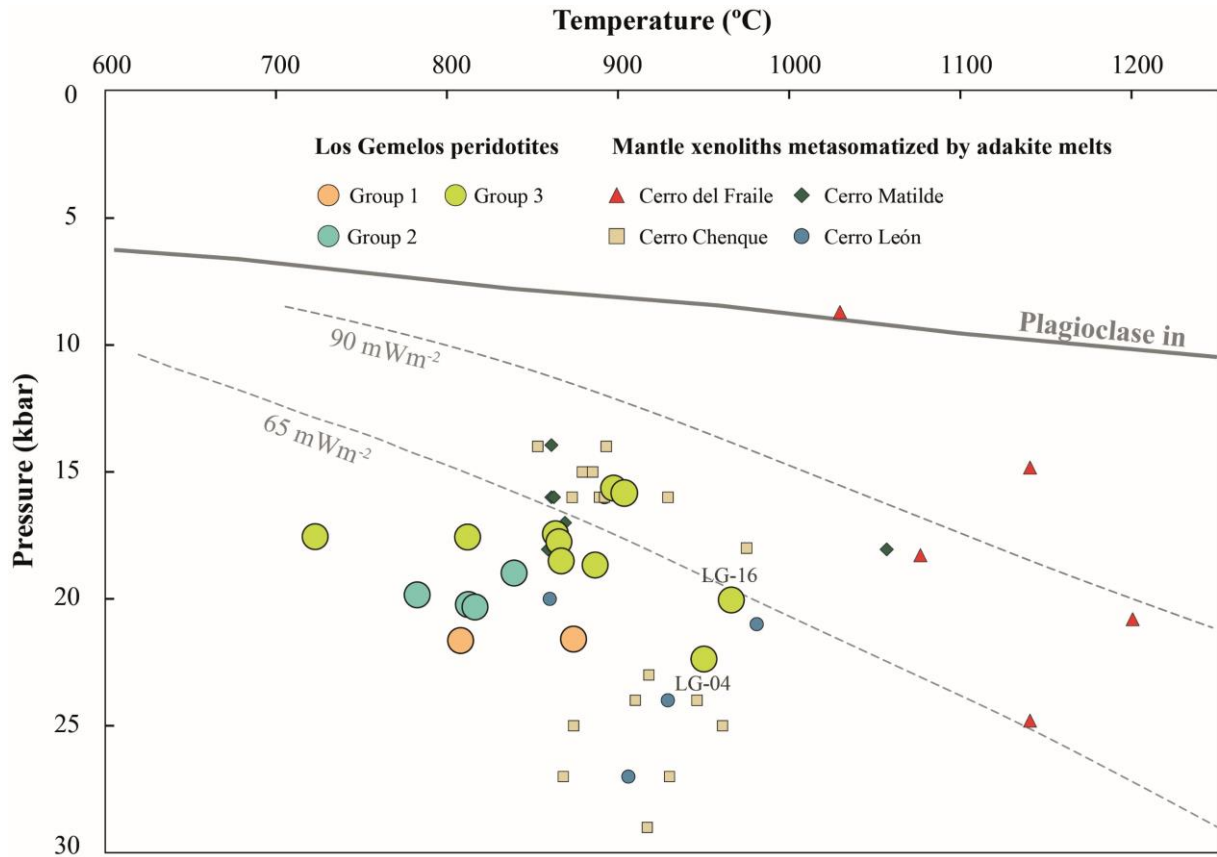


Figure S6. Pressure (15.6–22.4 kbar) and temperature (723–966 °C) estimates of Los Gemelos peridotites. In general, peridotites from Group 1 are deeper, whereas Group 2 represent the intermediate mantle section and Group 3 are the shallower. Samples LG-04 and LG-16 are highlighted because they show higher temperatures and pressures compared to the others from the same compositional group (Group 3). Pressure estimates were obtained by applying the two-pyroxenes model proposed by [Brey and Köhler \(1990\)](#). Temperature estimates were assumed as the average of two-pyroxenes ([Brey and Köhler, 1990](#); [Taylor, 1998](#)), Ca in orthopyroxene ([Brey and Köhler, 1990](#); [Nimis and Grütter, 2010](#)), and single-clinopyroxene ([Nimis and Taylor, 2000](#)) models. Pressure and temperature estimates for mantle xenoliths metasomatized by adakite melt metasomatism from Cerro del Fraile ([Kilian and Stern, 2002](#)) and Paso de Indios region (Cerro Chenque, Cerro Matilde, and Cerro León) ([Bertotto et al., 2021](#)) were plotted for comparison. The model geothermal gradients are from [Chapman \(1986\)](#).

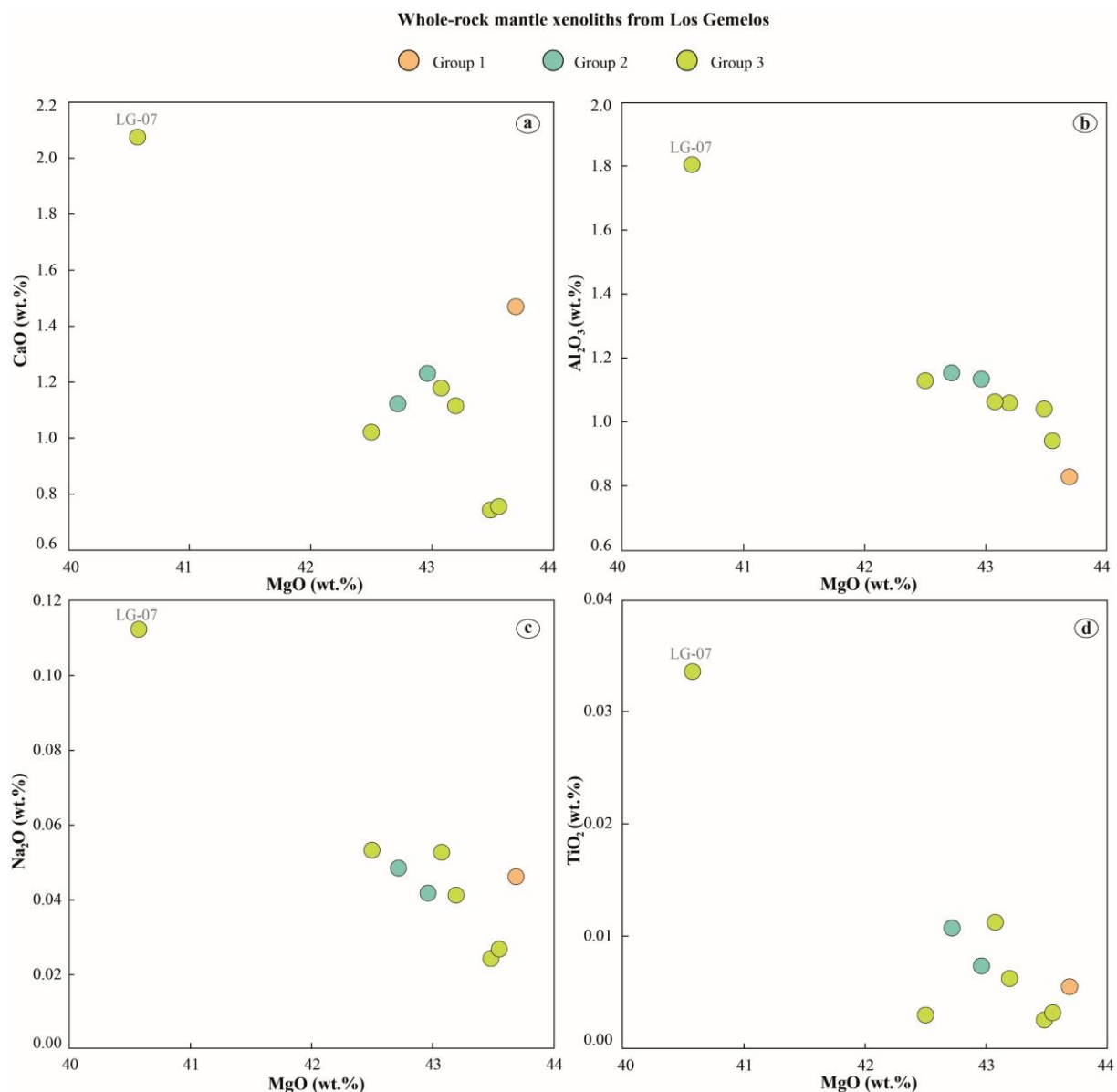


Figure S7. Selected major elements (CaO, Al_2O_3 , Na_2O , and TiO_2) vs. MgO of whole-rock peridotites from Los Gemelos. In general, our samples present typical features of partial melt residues as demonstrated by their high MgO coupled with low CaO, Al_2O_3 , Na_2O , and TiO_2 contents. Sample LG-07 is highlighted due to its distinct composition compared to the other xenoliths.

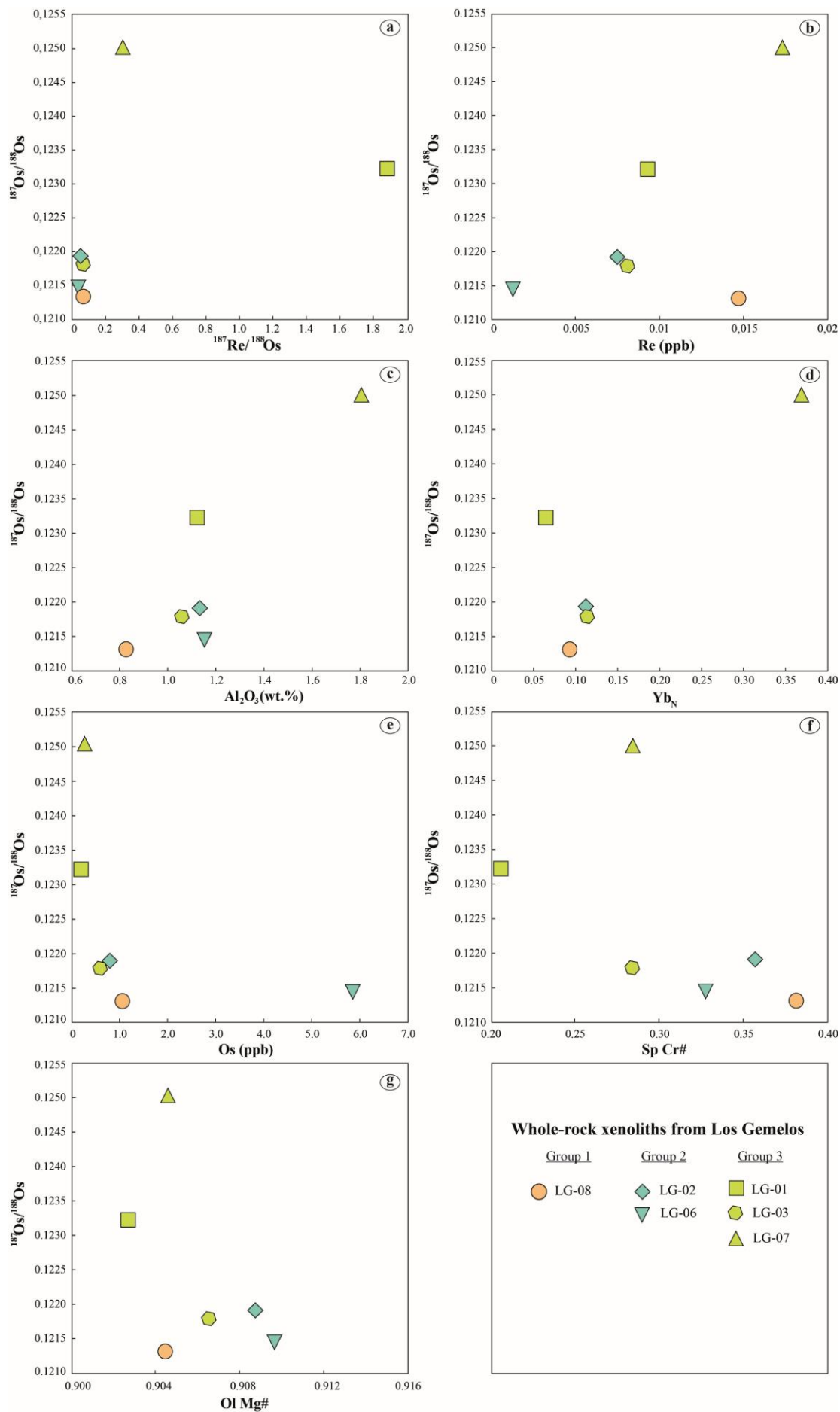


Figure S8. Whole-rock $^{187}\text{Os}/^{188}\text{Os}$ vs. $^{187}\text{Re}/^{188}\text{Os}$ (a), Re in ppb (b), Al_2O_3 in wt.% (c), Yb_N (d), Os in ppb (e), Sp Cr# (f), and Ol Mg# (g). Ytterbium is normalized (_N) to the primitive mantle ([Sun and McDonough, 1989](#)). $\text{Cr\#} = \text{Cr}/(\text{Cr}+\text{Al})$ and $\text{Mg\#} = \text{Mg}/(\text{Mg}+\text{Fe})$. Positive correlations are observed from (a) to (d), where in some samples (LG-01 and LG-08) plot out of the compositional trends. Negative correlations are defined from (e) to (g), with samples LG-06, LG-07, and LG-08 plotting out of the compositional trends.

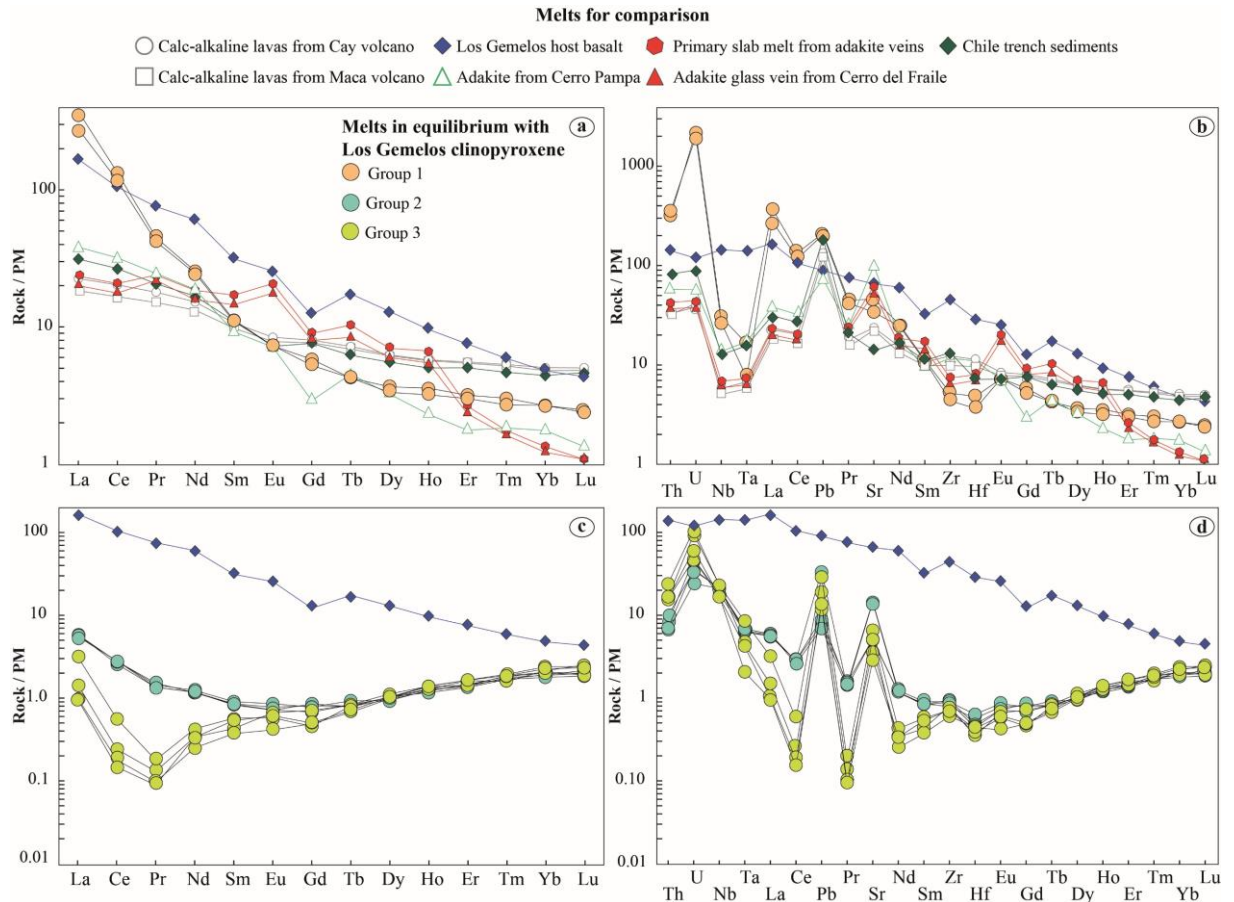


Figure S9. Potential melts in equilibrium with clinopyroxenes from Los Gemelos peridotites were calculated using the model proposed by [Ionov et al. \(2002\)](#), as well as their clinopyroxene/melt partition coefficients. The host basalt from Los Gemelos (average of three measurements; [Supplementary Table S1](#)), adakite glass vein and primary slab melt compositions from Cerro del Fraile ([Kilian and Stern, 2002](#)), the average of calc-alkaline lavas from Cay and Maca stratovolcanoes ([D'Orazio et al., 2003](#)), an adakite lava from Cerro Pampa ([Corgne et al., 2018](#)), and the average of Chile Trench sediments ([Kilian and Behrmann, 2003](#)) are plotted for comparison. Data are normalized to the primitive mantle ([Sun and McDonough, 1989](#)).

Table S1. Whole-rock geochemistry and K-Ar ages for Los Gemelos host lava.

Sample No.	Y566	Y570	Y571
Rock type	basanite	basanite	basanite
<i>Major elements (wt.%)</i>			
SiO₂	44,21	43,33	43,89
TiO₂	2,68	2,68	2,57
Al₂O₃	13,62	13,09	13,63
Fe₂O₃	12,97	13,26	12,24
MnO	0,20	0,20	0,21
MgO	9,37	10,04	8,66
CaO	9,27	9,24	9,06
Na₂O	5,17	4,60	5,65
K₂O	2,60	2,57	2,72
P₂O₅	1,26	1,28	1,25
Total	101,34	100,30	99,86
FeO*/MgO	1,24	1,19	1,27
Na₂O+K₂O	7,76	7,17	8,37
Mg#	58,9	60,0	58,3

<i>Compatible elements (ppm)</i>			
Cr	251,30	282,40	219,10
Ni	177,50	227,30	153,10

Table S1. Continued.

<i>Trace elements (ppm)</i>	Y566	Y570	Y571
Rb	47,50	43,60	49,30
Ba	734,90	695,00	767,90
Th	11,94	10,91	13,21
U	2,46	2,49	2,52
Nb	101,90	96,60	110,20
Ta	5,60	5,45	6,10
K	21566	21342	22570
La	109,31	101,79	134,73
Ce	183,36	163,38	213,96
Pb	6,45	5,30	7,29
Pr	20,16	18,77	23,48
Sr	1363,10	1239,40	1542,40
P	5503	5599	5451
Nd	79,11	75,46	91,80
Sm	13,42	13,02	15,32
Zr	500,00	467,90	557,20
Hf	8,57	8,62	9,49
Eu	4,18	3,98	4,70
Ti	16038	16044	15373
Gd	9,93	9,96	11,25
Tb	1,36	1,30	1,54
Dy	6,77	6,59	7,80
Y	28,40	26,90	29,90
Ho	1,18	1,16	1,35
Er	2,74	2,67	3,06
Tm	0,34	0,33	0,39
Yb	2,07	2,01	2,32
Lu	0,28	0,27	0,30

Table S1. Continued.

<i>PM-normalized values</i>	Y566	Y570	Y571
Rb	74,80	68,66	77,64
Ba	105,15	99,44	109,87
Th	140,49	128,33	155,42
U	117,15	118,39	120,11
Nb	142,92	135,48	154,56
Ta	136,61	133,02	148,71
K	86,26	85,37	90,28
La	159,11	148,16	196,12
Ce	103,30	92,04	120,54
Pb	90,89	74,62	102,64
Pr	73,05	68,01	85,06
Sr	64,60	58,74	73,10
P	57,93	58,94	57,38
Nd	58,43	55,73	67,80
Sm	30,22	29,33	34,50
Zr	44,64	41,78	49,75
Hf	27,74	27,89	30,70
Eu	24,87	23,71	27,96
Ti	12,34	12,34	11,83
Gd	16,66	16,71	18,88
Tb	12,58	12,05	14,26
Dy	9,18	8,95	10,58
Y	6,24	5,91	6,57
Ho	7,20	7,08	8,23
Er	5,70	5,57	6,37
Tm	4,60	4,40	5,30
Yb	4,19	4,07	4,71
Lu	3,76	3,59	4,04

<i>PM-normalized trace element ratios</i>	Y566	Y570	Y571
Ce/Yb	24,64	22,60	25,60
Nb/Nb*	1,05	1,02	1,01
Pb/Pb*	1,05	0,94	1,01

Table S1. Continued.

Sample No.	K (wt. %)			^{40}Ar rad (10^{-8} cm³STP/g)			$^{38}\text{Ar}/^{36}\text{Ar}$			Age (Ma)			Air Fraction (%)
Y566	2,038	±	0,061	3,73	±	0,20	0,18848	±	0,00051	0,47	±	0,03	83,89
Y570	2,163	±	0,065	4,50	±	0,23	0,18828	±	0,00057	0,54	±	0,03	90,25
Y571	2,249	±	0,067	3,00	±	0,16	0,18746	±	0,00068	0,34	±	0,02	84,12

Uncertainties are 1SD

Table S2. Representative (average) and individual major element contents (in wt.%) of olivine, orthopyroxene, clinopyroxene, spinel, phlogopite (LG-08 and LG-12), and adakite veins (LG-02, LG-06, LG-08, LG-12, LG-14, LG-15) from Los Gemelos mantle xenoliths.

Olivine																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Fo
LG-01	40,87	0,03	0,02	0,03	49,34	0,04	0,13	9,27	0,38	0,05	0,04	0,04	0,00	0,02	100,25	0,90	0,90
LG-02	41,37	0,04	0,01	0,04	49,18	0,02	0,12	8,44	0,39	0,00	0,01	0,01	0,00	0,02	99,66	0,91	0,91
LG-03	40,61	0,03	0,01	0,02	49,45	0,04	0,12	8,78	0,40	0,02	0,01	0,01	0,00	0,02	99,52	0,91	0,91
LG-04	41,39	0,04	0,01	0,02	48,92	0,03	0,13	8,39	0,40	0,01	0,01	0,01	0,00	0,02	99,36	0,91	0,91
LG-05	40,49	0,04	0,01	0,02	48,93	0,04	0,12	8,78	0,39	0,03	0,02	0,01	0,00	0,01	98,89	0,91	0,91
LG-06	40,74	0,07	0,01	0,01	50,35	0,03	0,13	8,51	0,41	0,01	0,01	0,00	0,00	0,01	100,29	0,91	0,91
LG-07	40,43	0,06	0,01	0,02	50,29	0,03	0,13	9,20	0,40	0,02	0,01	0,01	0,00	0,02	100,63	0,91	0,91
LG-08	40,60	0,04	0,02	0,05	49,07	0,04	0,10	9,01	0,37	0,04	0,01	0,01	0,00	0,01	99,37	0,91	0,91
LG-09	40,61	0,03	0,00	0,01	48,96	0,03	0,14	8,89	0,40	0,01	0,01	0,00	0,00	0,01	99,09	0,91	0,91
LG-10	40,71	0,03	0,01	0,05	48,59	0,04	0,14	8,96	0,38	0,01	0,00	0,00	0,00	0,01	98,94	0,91	0,90
LG-11	41,51	0,06	0,01	0,01	48,64	0,04	0,11	8,87	0,40	0,01	0,01	0,00	0,00	0,01	99,68	0,91	0,91
LG-12	40,47	0,01	0,01	0,03	49,35	0,03	0,16	8,78	0,38	0,01	0,01	0,00	0,00	0,02	99,25	0,91	0,91
LG-13	40,30	0,05	0,02	0,02	50,89	0,02	0,11	8,41	0,43	0,01	0,00	0,00	0,00	0,01	100,27	0,92	0,91
LG-14	40,52	0,04	0,01	0,02	50,46	0,03	0,11	8,27	0,41	0,01	0,01	0,00	0,00	0,02	99,89	0,92	0,91
LG-15	41,53	0,03	0,01	0,03	49,48	0,03	0,13	8,35	0,43	0,01	0,01	0,01	0,00	0,01	100,05	0,91	0,91
LG-16	41,32	0,05	0,01	0,02	49,49	0,03	0,13	8,75	0,40	0,02	0,01	0,01	0,00	0,02	100,25	0,91	0,91

Table S2. Continued.

Orthopyroxene																			
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Ens	Fs	Wo
LG-01	56,49	0,08	2,95	0,43	33,58	0,53	0,11	5,85	0,10	0,12	0,05	0,05	0,00	0,03	100,36	0,91	90,01	8,79	1,03
LG-02	57,14	0,04	2,36	0,44	33,58	0,45	0,12	5,36	0,08	0,06	0,02	0,01	0,00	0,03	99,68	0,92	90,81	8,13	0,88
LG-03	56,87	0,03	2,57	0,38	33,25	0,49	0,15	5,64	0,09	0,05	0,02	0,01	0,00	0,02	99,57	0,91	90,21	8,59	0,96
LG-04	57,59	0,06	1,83	0,37	33,62	0,39	0,14	5,47	0,09	0,07	0,01	0,01	0,00	0,02	99,66	0,92	90,74	8,28	0,76
LG-05	56,21	0,07	2,57	0,39	33,56	0,49	0,13	5,57	0,09	0,05	0,02	0,01	0,00	0,03	99,18	0,91	90,43	8,42	0,95
LG-06	56,53	0,04	2,28	0,45	34,75	0,42	0,13	5,46	0,09	0,04	0,01	0,00	0,00	0,02	100,24	0,92	90,99	8,02	0,80
LG-07	55,70	0,04	2,57	0,40	34,60	0,49	0,15	5,83	0,10	0,05	0,03	0,02	0,00	0,02	100,00	0,91	90,31	8,55	0,93
LG-08	57,50	0,06	1,88	0,32	33,43	0,34	0,13	5,68	0,08	0,06	0,02	0,02	0,00	0,02	99,55	0,91	90,51	8,62	0,67
LG-09	56,58	0,04	2,45	0,41	33,48	0,48	0,15	5,67	0,10	0,02	0,01	0,00	0,00	0,02	99,41	0,91	90,27	8,57	0,94
LG-10	56,25	0,07	2,97	0,43	33,11	0,51	0,13	5,72	0,10	0,05	0,01	0,01	0,00	0,02	99,37	0,91	90,07	8,73	1,01
LG-11	56,82	0,06	2,45	0,40	33,18	0,49	0,14	5,55	0,10	0,02	0,00	0,00	0,00	0,03	99,24	0,91	90,34	8,48	0,97
LG-12	56,95	0,06	1,99	0,39	33,68	0,42	0,12	5,58	0,08	0,04	0,01	0,00	0,00	0,02	99,33	0,91	90,59	8,42	0,81
LG-13	56,51	0,05	2,60	0,44	34,52	0,41	0,13	5,44	0,08	0,07	0,01	0,01	0,00	0,02	100,29	0,92	90,98	8,05	0,77
LG-14	56,44	0,01	2,20	0,45	35,44	0,48	0,14	5,27	0,10	0,04	0,01	0,00	0,00	0,02	100,61	0,92	91,28	7,62	0,89
LG-15	57,45	0,04	2,23	0,45	33,51	0,44	0,14	5,31	0,11	0,02	0,00	0,00	0,00	0,02	99,73	0,92	90,85	8,08	0,86
LG-16	57,48	0,03	2,20	0,40	33,38	0,49	0,12	5,59	0,10	0,03	0,00	0,01	0,00	0,03	99,87	0,91	90,35	8,50	0,96

Table S2. Continued.

Clinopyroxene																				
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Ens	Fs	Wo	Ca/Al
LG-01	53,50	0,14	4,08	0,93	15,58	21,63	0,08	2,26	0,05	1,10	0,04	0,02	0,00	0,05	99,46	0,92	48,01	3,91	47,94	4,83
LG-02	54,50	0,04	2,61	0,89	16,46	22,96	0,07	1,83	0,04	0,68	0,01	0,01	0,00	0,03	100,13	0,94	48,37	3,02	48,50	8,05
LG-03	53,93	0,07	2,71	0,73	16,58	22,84	0,08	2,04	0,06	0,58	0,02	0,01	0,00	0,03	99,67	0,94	48,49	3,34	48,04	7,72
LG-04	54,88	0,08	3,77	1,54	15,69	21,44	0,09	2,03	0,04	0,90	0,01	0,01	0,00	0,05	100,52	0,93	48,60	3,53	47,73	5,18
LG-05	53,91	0,06	2,64	0,72	16,59	22,81	0,08	2,00	0,06	0,57	0,03	0,02	0,00	0,04	99,52	0,94	48,57	3,28	48,02	7,91
LG-06	53,78	0,06	2,50	0,83	16,99	23,28	0,06	1,88	0,05	0,67	0,01	0,01	0,00	0,04	100,18	0,94	48,79	3,03	48,08	8,50
LG-07	54,00	0,05	2,63	0,79	16,42	23,32	0,07	2,11	0,04	0,57	0,02	0,01	0,00	0,04	100,07	0,93	47,72	3,45	48,72	8,07
LG-08	54,54	0,15	2,56	0,91	15,93	22,02	0,06	1,92	0,03	0,81	0,03	0,02	0,00	0,04	99,02	0,94	48,44	3,28	48,17	7,86
LG-09	53,62	0,08	2,53	0,73	16,62	22,67	0,09	2,04	0,06	0,58	0,00	0,01	0,00	0,03	99,07	0,94	48,72	3,36	47,77	8,16
LG-10	53,51	0,15	3,93	0,92	15,73	21,55	0,10	2,23	0,07	1,10	0,01	0,01	0,00	0,04	99,33	0,93	48,35	3,85	47,63	5,01
LG-11	53,90	0,09	2,59	0,77	16,50	22,58	0,07	2,02	0,06	0,57	0,01	0,01	0,00	0,04	99,19	0,94	48,66	3,34	47,88	7,97
LG-12	54,14	0,16	2,64	1,01	16,14	22,47	0,07	1,92	0,04	0,97	0,01	0,01	0,00	0,04	99,62	0,94	48,31	3,22	48,35	7,84
LG-13	53,91	0,07	2,62	1,13	16,15	23,28	0,08	2,05	0,05	0,84	0,01	0,00	0,00	0,04	100,24	0,93	47,39	3,37	49,11	8,16
LG-14	53,73	0,07	2,36	0,90	17,36	23,09	0,07	1,80	0,06	0,70	0,01	0,00	0,00	0,03	100,18	0,95	49,58	2,88	47,42	8,94
LG-15	54,43	0,06	2,44	0,89	16,49	23,08	0,06	1,82	0,06	0,69	0,01	0,00	0,00	0,03	100,06	0,94	48,30	2,99	48,60	8,67
LG-16	54,79	0,06	3,46	1,71	15,52	21,95	0,09	2,05	0,05	0,57	0,01	0,01	0,00	0,03	100,29	0,93	47,76	3,54	48,55	5,77

Table S2. Continued.

Spinel																						
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Cr#	Mg#	Spinel	Chromite	Magnetite	Fe ₂ O ₃	FeO
LG-01	0,01	0,06	48,60	18,59	19,59	0,01	0,12	11,50	0,31	0,05	0,02	0,03	0,00	0,12	99,28	0,20	0,79	0,79	0,20	0,03	2,74	9,03
LG-02	0,01	0,07	43,45	25,44	18,00	0,01	0,13	11,57	0,25	0,01	0,01	0,00	0,00	0,10	99,14	0,28	0,75	0,74	0,28	0,01	0,94	10,72
LG-02	0,00	0,02	37,87	31,35	17,45	0,01	0,17	12,53	0,19	0,01	0,01	0,00	0,00	0,12	99,92	0,36	0,74	0,73	0,35	0,02	1,82	10,90
LG-03	0,01	0,07	42,70	25,31	18,49	0,01	0,15	12,07	0,27	0,01	0,01	0,00	0,00	0,13	99,47	0,28	0,77	0,76	0,28	0,03	2,42	9,89
LG-04	0,01	0,03	33,48	35,52	15,26	0,01	0,18	14,73	0,15	0,02	0,01	0,01	0,00	0,12	99,67	0,42	0,67	0,66	0,41	0,01	1,35	13,51
LG-05	0,01	0,07	42,65	25,50	18,56	0,01	0,17	12,59	0,23	0,02	0,00	0,00	0,00	0,12	100,22	0,29	0,77	0,76	0,28	0,03	2,84	10,04
LG-06	0,01	0,04	39,85	28,95	17,53	0,03	0,14	12,37	0,23	0,02	0,01	0,00	0,00	0,13	99,45	0,33	0,74	0,73	0,32	0,02	1,58	10,94
LG-07	0,01	0,05	43,02	25,45	17,88	0,01	0,16	12,54	0,25	0,01	0,01	0,01	0,00	0,12	99,70	0,28	0,74	0,74	0,28	0,02	1,77	10,95
LG-08	0,01	0,12	35,59	32,71	16,94	0,02	0,17	13,43	0,17	0,03	0,01	0,01	0,00	0,17	99,61	0,38	0,73	0,72	0,37	0,03	2,38	11,28
LG-09	0,00	0,05	42,67	25,40	18,80	0,01	0,14	11,92	0,27	0,01	0,01	0,00	0,00	0,12	99,68	0,29	0,78	0,77	0,28	0,03	2,72	9,47
LG-10	0,00	0,08	48,40	19,02	19,82	0,01	0,15	11,19	0,33	0,01	0,01	0,01	0,00	0,09	99,40	0,21	0,80	0,80	0,20	0,03	2,77	8,69
LG-11	0,00	0,06	42,71	25,21	18,58	0,01	0,16	11,97	0,26	0,01	0,01	0,01	0,00	0,12	99,37	0,28	0,77	0,77	0,28	0,03	2,50	9,72
LG-12	0,00	0,16	35,97	32,96	16,70	0,01	0,19	13,04	0,17	0,01	0,00	0,01	0,00	0,18	99,55	0,38	0,72	0,71	0,37	0,02	1,45	11,74
LG-13	0,01	0,07	42,70	25,74	17,68	0,01	0,19	12,60	0,16	0,01	0,01	0,00	0,00	0,12	99,45	0,29	0,74	0,73	0,28	0,02	1,53	11,22
LG-14	0,00	0,08	35,73	32,54	17,71	0,02	0,16	12,34	0,22	0,01	0,01	0,01	0,00	0,12	99,22	0,38	0,76	0,75	0,37	0,03	2,66	9,95
LG-15	0,01	0,06	40,79	27,67	17,57	0,02	0,14	11,96	0,23	0,01	0,01	0,01	0,00	0,11	98,72	0,31	0,74	0,74	0,31	0,01	1,26	10,82
LG-16	0,02	0,04	39,82	27,54	18,58	0,01	0,14	12,14	0,26	0,01	0,01	0,01	0,00	0,13	99,02	0,32	0,78	0,78	0,31	0,04	3,34	9,13

Table S2. Continued.

Phlogopite (euohedral)																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	
LG-08	39,91	2,11	15,18	1,28	22,12	0,04	0,04	3,24	0,21	1,05	8,79	0,24	0,01	0,07	94,29	0,92	
LG-12	38,45	2,21	15,08	1,55	22,50	0,03	0,04	3,27	0,24	1,00	8,82	0,23	0,12	0,06	93,61	0,92	
Phlogopite (inclusion in spinel)																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	
LG-08	39,61	2,31	15,38	1,04	22,60	0,11	0,03	3,35	0,15	1,00	8,85	0,19	0,05	0,04	94,70	0,92	
LG-12	38,69	2,28	15,35	1,42	22,64	0,04	0,02	3,40	0,18	0,99	8,96	0,18	0,18	0,06	94,40	0,92	
K-rich alkaline melt																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	
LG-02	60,38	2,35	16,52	0,03	2,35	5,73	0,08	3,46	0,02	4,54	2,84	0,07	0,01	0,05	98,43	0,57	
LG-06	48,75	2,42	16,56	0,02	4,59	12,14	0,08	5,58	0,02	4,33	3,14	0,10	0,03	0,05	97,81	0,62	
LG-08	50,19	2,20	16,65	0,04	4,85	7,08	0,13	7,18	0,03	5,32	4,27	0,11	0,06	0,06	98,16	0,57	
LG-12	50,93	2,42	17,15	0,01	4,66	5,82	0,04	7,12	0,03	5,20	4,37	0,24	0,03	0,03	98,06	0,56	
LG-14a	61,02	2,47	16,23	0,01	2,37	4,75	0,05	2,68	0,03	4,52	3,27	0,07	0,01	0,04	98,06	0,64	
LG-14b	50,47	2,74	17,62	0,02	4,43	7,76	0,12	6,03	0,03	4,91	3,39	0,09	0,07	0,05	97,72	0,59	
LG-15a	48,89	2,91	17,14	0,02	4,43	11,86	0,11	5,11	0,02	4,14	3,23	0,10	0,04	0,10	98,10	0,63	
LG-15b	61,29	2,42	15,78	0,02	2,26	5,45	0,05	2,70	0,01	4,40	3,65	0,06	0,00	0,06	98,15	0,62	
Metasomatic minerals in veins																	
Type 2 olivine																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	
LG-02	40,20	0,03	0,03	0,04	45,56	0,25	0,22	13,18	0,12	0,02	0,02	0,01	0,00	0,01	99,68	0,86	
LG-14	40,32	0,12	0,02	0,07	47,00	0,23	0,20	11,60	0,13	0,03	0,01	0,01	0,00	0,02	99,74	0,88	
Type 2 orthopyroxene																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	
LG-14	56,59	0,07	2,38	0,51	34,89	0,45	0,14	4,77	0,09	0,02	0,01	0,01	0,01	0,01	99,94	0,93	
LG-15	56,42	0,07	2,16	0,02	34,50	0,50	0,13	5,12	0,11	0,05	0,03	0,01	0,16	0,02	99,29	0,92	
Type 2 clinopyroxene																	
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Ca/Al
LG-02	52,90	0,23	2,41	0,87	18,43	21,44	0,12	2,73	0,05	0,71	0,07	0,01	0,00	0,08	100,07	0,92	8,16
LG-14	52,45	0,10	2,60	1,01	18,91	21,31	0,20	2,60	0,05	1,00	0,03	0,00	0,00	0,04	100,30	0,93	7,58

Table S2. Continued.

Mineral inclusions in spinel																			
Olivine																			
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#			
LG-02	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00			
LG-14	SiO2	TiO2	Al2O3	Cr2O3	MgO	CaO	MnO	FeO	NiO	Na2O	K2O	Cl	F	V2O3	Total	Mg#			
Orthopyroxene																			
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#			
LG-08	56,64	0,13	1,67	0,01	34,61	0,36	0,12	5,52	0,07	0,05	0,01	0,01	0,04	0,02	99,24	0,92			
LG-12	56,84	0,03	1,68	0,03	34,58	0,38	0,11	5,62	0,11	0,03	0,01	0,00	0,00	0,03	99,45	0,92			
Clinopyroxene																			
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Ca/Al		
LG-08	53,87	0,17	2,38	0,04	16,66	22,93	0,07	1,91	0,05	0,80	0,02	0,01	0,07	0,05	99,02	0,94	8,75		
LG-12	53,78	0,12	2,35	0,00	16,55	22,91	0,05	1,91	0,06	0,88	0,01	0,00	0,00	0,02	98,64	0,94	8,85		
Cpx exsolution in Opx																			
Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	MgO	CaO	MnO	FeO	NiO	Na ₂ O	K ₂ O	Cl	F	V ₂ O ₃	Total	Mg#	Ens	Fs	Wo
LG-02	54,60	0,00	2,77	0,90	16,06	23,05	0,09	1,90	0,04	0,74	0,02	0,02	0,00	0,05	100,23	0,94	47,58	3,16	49,10

Table S3. Representative (average) and individual trace element contents (in ppm) of clinopyroxene and phlogopite (LG-08 and LG-12) of Los Gemelos mantle xenoliths.

Clinopyroxene																														
Sample	Rb	Ba	W	Th	U	Nb	Ta	K	La	Ce	Pb	Pr	Mo	Sr	Nd	Sm	Zr	Hf	Eu	Ti	Gd	Tb	Dy	Li	Y	Ho	Er	Tm	Yb	Lu
LG-02	b.d.l	0,03	b.d.l	0,01	0,01	0,10	0,01	6,34	0,21	0,43	0,17	0,05	0,08	13,97	0,30	0,11	1,15	0,04	0,05	153,60	0,19	0,04	0,34	0,80	2,21	0,08	0,28	0,05	0,37	0,06
LG-03	b.d.l	0,06	b.d.l	0,02	0,02	0,13	0,00	8,86	0,05	0,04	0,10	0,01	0,09	10,28	0,09	0,05	1,16	0,02	0,04	216,53	0,17	0,04	0,37	1,02	2,70	0,10	0,35	0,06	0,50	0,07
LG-06	b.d.l	0,06	b.d.l	0,01	0,01	0,12	0,01	6,46	0,22	0,43	0,04	0,06	0,04	13,95	0,29	0,11	1,37	0,03	0,04	171,25	0,16	0,04	0,33	1,01	2,52	0,10	0,30	0,06	0,43	0,06
LG-08	0,06	0,49	0,55	0,34	0,46	0,17	0,01	7,40	14,01	20,65	1,07	1,75	0,05	123,93	6,48	1,44	7,56	0,30	0,43	876,18	1,40	0,20	1,20	2,33	6,18	0,25	0,67	0,09	0,56	0,07
LG-09	b.d.l	0,22	b.d.l	0,03	0,02	0,13	0,01	9,66	0,12	0,09	0,15	0,01	0,08	18,31	0,11	0,07	1,04	0,02	0,03	181,22	0,11	0,04	0,34	0,88	2,49	0,09	0,34	0,05	0,45	0,07
LG-11	b.d.l	0,08	0,02	0,02	0,01	0,09	0,00	7,28	0,04	0,03	0,06	0,00	0,11	14,63	0,06	0,05	0,88	0,03	0,02	172,66	0,12	0,03	0,32	1,00	2,35	0,09	0,31	0,06	0,42	0,06
LG-12	0,03	2,21	0,30	0,37	0,41	0,15	0,01	5,78	10,41	18,03	1,03	1,64	0,07	91,47	6,23	1,40	6,34	0,23	0,44	750,44	1,26	0,20	1,09	2,42	5,72	0,23	0,63	0,08	0,56	0,07
LG-14	b.d.l	0,10	0,17	0,01	0,00	0,12	0,01	9,34	0,21	0,43	0,05	0,06	0,06	39,61	0,31	0,12	1,31	0,04	0,05	170,70	0,19	0,04	0,31	1,00	2,27	0,09	0,29	0,05	0,40	0,06
LG-15	b.d.l	0,02	b.d.l	0,01	0,01	0,11	0,01	5,54	0,22	0,40	0,04	0,05	0,04	38,87	0,30	0,11	1,30	0,03	0,04	162,98	0,20	0,04	0,32	1,01	2,46	0,09	0,30	0,05	0,40	0,06
LG-16	b.d.l	0,04	b.d.l	0,02	0,01	0,09	0,00	8,80	0,04	0,02	0,07	0,00	0,07	7,76	0,08	0,07	1,04	0,03	0,04	183,72	0,12	0,03	0,34	0,99	2,59	0,10	0,34	0,06	0,48	0,07

Phlogopite (euhedral)																															
Sample	Cs	Rb	Ba	W	Th	U	Nb	Ta	K	La	Ce	Pb	Pr	Mo	Sr	Nd	Sm	Zr	Hf	Eu	Ti	Gd	Tb	Dy	Li	Y	Ho	Er	Tm	Yb	Lu
LG-08	2,27	143,99	1181,81	1,78	0,05	0,21	29,67	0,62	50836,35	1,74	1,56	8,44	0,16	0,30	584,09	0,57	0,11	1,88	0,05	0,13	12785,29	0,09	0,01	0,05	2,00	0,42	0,01	0,04	0,00	0,03	0,01
LG-12	1,78	139,66	808,34	1,86	0,02	0,20	27,15	0,50	48179,27	0,64	0,87	8,45	0,05	0,38	678,48	0,17	0,02	1,72	0,05	0,10	13309,07	0,03	0,01	0,02	0,93	0,17	0,01	0,02	0,00	0,01	0,00

Clinopyroxene																														
Sample	Rb	Ba	W	Th	U	Nb	Ta	K	La	Ce	Pb	Pr	Mo	Sr	Nd	Sm	Zr	Hf	Eu	Ti	Gd	Tb	Dy	Li	Y	Ho	Er	Tm	Yb	Lu
LG-02	b.d.l	0,03	b.d.l	0,01	0,01	0,10	0,01	6,34	0,21	0,43	0,17	0,05	0,08	13,97	0,30	0,11	1,15	0,04	0,05	153,60	0,19	0,04	0,34	0,80	2,21	0,08	0,28	0,05	0,37	0,06
LG-03	b.d.l	0,06	b.d.l	0,02	0,02	0,13	0,00	8,86	0,05	0,04	0,10	0,01	0,09	10,28	0,09	0,05	1,16	0,02	0,04	216,53	0,17	0,04	0,37	1,02	2,70	0,10	0,35	0,06	0,50	0,07
LG-06	b.d.l	0,06	b.d.l	0,01	0,01	0,12	0,01	6,46	0,22	0,43	0,04	0,06	0,04	13,95	0,29	0,11	1,37	0,03	0,04	171,25	0,16	0,04	0,33	1,01	2,52	0,10	0,30	0,06	0,43	0,06
LG-08	0,06	0,49	0,55	0,34	0,46	0,17	0,01	7,40	14,01	20,65	1,07	1,75	0,05	123,93	6,48	1,44	7,56	0,30	0,43	876,18	1,40	0,20	1,20	2,33	6,18	0,25	0,67	0,09	0,56	0,07
LG-09	b.d.l	0,22	b.d.l	0,03	0,02	0,13	0,01	9,66	0,12	0,09	0,15	0,01	0,08	18,31	0,11	0,07	1,04	0,02	0,03	181,22	0,11	0,04	0,34	0,88	2,49	0,09	0,34	0,05	0,45	0,07
LG-11	b.d.l	0,08	0,02	0,02	0,01	0,09	0,00	7,28	0,04	0,03	0,06	0,00	0,11	14,63	0,06	0,05	0,88	0,03	0,02	172,66	0,12	0,03	0,32	1,00	2,35	0,09	0,31	0,06	0,42	0,06

UnB IG PPG Geologia																														
LG-12	0,03	2,21	0,30	0,37	0,41	0,15	0,01	5,78	10,41	18,03	1,03	1,64	0,07	91,47	6,23	1,40	6,34	0,23	0,44	750,44	1,26	0,20	1,09	2,42	5,72	0,23	0,63	0,08	0,56	0,07
LG-14	b.d.l	0,10	0,17	0,01	0,00	0,12	0,01	9,34	0,21	0,43	0,05	0,06	0,06	39,61	0,31	0,12	1,31	0,04	0,05	170,70	0,19	0,04	0,31	1,00	2,27	0,09	0,29	0,05	0,40	0,06
LG-15	b.d.l	0,02	b.d.l	0,01	0,01	0,11	0,01	5,54	0,22	0,40	0,04	0,05	0,04	38,87	0,30	0,11	1,30	0,03	0,04	162,98	0,20	0,04	0,32	1,01	2,46	0,09	0,30	0,05	0,40	0,06
LG-16	b.d.l	0,04	b.d.l	0,02	0,01	0,09	0,00	8,80	0,04	0,02	0,07	0,00	0,07	7,76	0,08	0,07	1,04	0,03	0,04	183,72	0,12	0,03	0,34	0,99	2,59	0,10	0,34	0,06	0,48	0,07

Table S2.
Continued.

Phlogopite (euhedral)																														
Sample	Cs	Rb	Ba	W	Th	U	Nb	Ta	K	La	Ce	Pb	Pr	Mo	Sr	Nd	Sm	Zr	Hf	Eu	Ti	Gd	Tb	Dy	Li	Y	Ho	Er	Tm	Yb
LG-08	2,27	143,99	1181,81	1,78	0,05	0,21	29,67	0,62	50836,35	1,74	1,56	8,44	0,16	0,30	584,09	0,57	0,11	1,88	0,05	0,13	12785,29	0,09	0,01	0,05	2,00	0,42	0,01	0,04	0,00	0,03

Table S4. Whole-rock major and trace element contents of Los Gemelos mantle xenoliths.

Major elements (wt.%)									
Samples	LG-01	LG-02	LG-03	LG-05	LG-06	LG-07	LG-08	LG-10	LG-11
SiO₂	45,40	45,99	45,95	45,48	46,49	46,44	44,89	45,75	45,88
Al₂O₃	1,13	1,13	1,06	1,06	1,15	1,80	0,83	1,04	0,94
Fe₂O₃ (t)	7,95	8,17	7,70	8,22	7,73	8,37	8,11	7,85	7,86
MnO	0,12	0,12	0,12	0,12	0,12	0,12	0,12	0,12	0,12
MgO	42,50	42,96	43,19	43,08	42,72	40,57	43,69	43,48	43,55
CaO	1,02	1,23	1,11	1,18	1,12	2,07	1,47	0,74	0,75
Na₂O	0,05	0,04	0,04	0,05	0,05	0,11	0,05	0,02	0,03
K₂O	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00
TiO₂	0,00	0,01	0,01	0,01	0,01	0,03	0,01	0,00	0,00
P₂O₅	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00
#Mg	0,91	0,91	0,92	0,91	0,92	0,91	0,91	0,92	0,92
Total	98,18	99,66	99,18	99,21	99,38	99,54	99,16	99,01	99,13
LOI	-0,12	-0,53	-0,28	-0,13	-0,31	-0,13	-0,29	0,08	0,30

Trace elements (ppm)									
Samples	LG-01	LG-02	LG-03	LG-05	LG-06	LG-07	LG-08	LG-10	LG-11
Cs	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00
Rb	0,17	0,04	0,03	0,06	0,06	0,04	0,08	0,06	0,03
Ba	4,63	0,72	0,68	1,36	1,43	0,96	1,26	1,84	0,76
Th	0,05	0,01	0,01	0,02	0,01	0,01	0,02	0,02	0,01
U	0,01	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00
Nb	0,52	0,08	0,09	0,16	0,14	0,14	0,16	0,16	0,10
Ta	0,03	0,00	0,00	0,01	0,01	0,01	0,01	0,01	0,01
La	0,38	0,08	0,09	0,16	0,13	0,08	0,18	0,15	0,12
Ce	0,63	0,13	0,13	0,27	0,22	0,14	0,30	0,25	0,16
Pb	0,04	0,01	0,01	0,04	0,01	0,02	0,06	0,02	0,05
Pr	0,07	0,02	0,02	0,03	0,03	0,02	0,03	0,03	0,02
Sr	4,61	0,95	0,85	1,58	1,45	0,93	1,85	1,82	0,92
Nd	0,24	0,06	0,08	0,12	0,11	0,07	0,13	0,10	0,09
Sm	0,04	0,01	0,02	0,03	0,03	0,04	0,03	0,02	0,02
Zr	1,60	0,24	0,34	0,52	0,45	0,38	0,46	0,31	0,28
Hf	0,03	0,01	0,01	0,02	0,01	0,01	0,01	0,01	0,01
Eu	0,01	0,00	0,01	0,01	0,01	0,02	0,01	0,01	0,01
Ti	64,31	35,51	57,54	97,24	85,79	225,63	50,50	34,64	36,24
Gd	0,03	0,02	0,03	0,06	0,05	0,09	0,02	0,01	0,02
Tb	0,00	0,00	0,01	0,01	0,01	0,02	0,00	0,00	0,00
Dy	0,02	0,03	0,05	0,09	0,07	0,17	0,03	0,01	0,02
Y	0,14	0,24	0,33	0,60	0,48	1,19	0,24	0,11	0,15
Ho	0,00	0,01	0,01	0,02	0,02	0,05	0,01	0,00	0,01
Er	0,02	0,03	0,04	0,07	0,06	0,15	0,03	0,01	0,02
Tm	0,00	0,01	0,01	0,01	0,01	0,03	0,01	0,00	0,00
Yb	0,03	0,06	0,06	0,09	0,07	0,18	0,05	0,03	0,03
Lu	0,01	0,01	0,01	0,01	0,01	0,03	0,01	0,01	0,01

Table S4. Continued.

PM-normalized trace elements									
Samples	LG-01	LG-02	LG-03	LG-05	LG-06	LG-07	LG-08	LG-10	LG-11
Cs	0,44	0,08	0,09	0,21	0,13	0,11	0,16	0,13	0,07
Rb	0,26	0,06	0,05	0,10	0,09	0,07	0,12	0,10	0,05
Ba	0,66	0,10	0,10	0,19	0,21	0,14	0,18	0,26	0,11
Th	0,61	0,10	0,11	0,19	0,18	0,13	0,19	0,23	0,12
U	0,62	0,11	0,12	0,23	0,21	0,15	0,21	0,23	0,12
Nb	0,73	0,11	0,13	0,23	0,19	0,19	0,22	0,22	0,15
Ta	0,76	0,11	0,11	0,20	0,20	0,14	0,20	0,24	0,14
La	0,55	0,11	0,13	0,23	0,19	0,11	0,26	0,22	0,17
Ce	0,35	0,07	0,08	0,15	0,12	0,08	0,17	0,14	0,09
Pb	0,56	0,21	0,17	0,60	0,13	0,22	0,78	0,30	0,68
Pr	0,25	0,06	0,07	0,11	0,10	0,06	0,12	0,10	0,08
Sr	0,22	0,05	0,04	0,07	0,07	0,04	0,09	0,09	0,04
Nd	0,18	0,04	0,06	0,09	0,08	0,05	0,10	0,08	0,06
Sm	0,09	0,03	0,05	0,07	0,06	0,08	0,06	0,04	0,04
Zr	0,14	0,02	0,03	0,05	0,04	0,03	0,04	0,03	0,03
Hf	0,10	0,02	0,03	0,05	0,05	0,05	0,04	0,03	0,02
Eu	0,07	0,03	0,04	0,08	0,06	0,10	0,05	0,03	0,03
Ti	0,05	0,03	0,04	0,07	0,07	0,17	0,04	0,03	0,03
Gd	0,05	0,03	0,05	0,10	0,08	0,16	0,04	0,02	0,03
Tb	0,04	0,03	0,06	0,12	0,09	0,20	0,04	0,02	0,03
Dy	0,03	0,04	0,06	0,13	0,10	0,24	0,05	0,02	0,03
Y	0,03	0,05	0,07	0,13	0,11	0,26	0,05	0,02	0,03
Ho	0,03	0,05	0,08	0,14	0,11	0,28	0,05	0,02	0,03
Er	0,03	0,07	0,09	0,15	0,12	0,31	0,06	0,03	0,04
Tm	0,05	0,09	0,10	0,16	0,13	0,34	0,08	0,04	0,06
Yb	0,06	0,11	0,11	0,17	0,15	0,37	0,09	0,06	0,07
Lu	0,09	0,15	0,14	0,19	0,17	0,41	0,11	0,09	0,10

PM-normalized ratios									
Samples	LG-01	LG-02	LG-03	LG-05	LG-06	LG-07	LG-08	LG-10	LG-11
La/Sm	6,31	3,88	2,94	3,18	3,08	1,34	4,48	5,32	4,76
Sm/Yb	1,35	0,26	0,40	0,42	0,41	0,23	0,61	0,65	0,51
Dy/Yb	0,47	0,34	0,57	0,72	0,66	0,64	0,51	0,32	0,36
Ce/Yb	5,46	0,67	0,67	0,86	0,81	0,22	1,80	2,27	1,32
Gd/Yb	0,75	0,25	0,47	0,55	0,52	0,42	0,44	0,39	0,43

Table S5. Whole-rock Sr-Nd isotopic ratios from Los Gemelos mantle xenoliths.

Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	2σ	$^{84}\text{Sr}/^{86}\text{Sr}$	2σ	$^{143}\text{Nd}/^{144}\text{Nd}$	2σ	$^{145}\text{Nd}/^{144}\text{Nd}$	2σ	$^{150}\text{Nd}/^{144}\text{Nd}$	2σ
LG-01	0,704174	0,000011	0,056493	0,000005	0,512723	0,000009	0,348398	0,000006	0,236412	0,000011
LG-03	0,704608	0,000011	0,056502	0,000007	0,512752	0,000024	0,348396	0,000014	0,236436	0,000024
LG-05	0,704378	0,000010	0,056499	0,000007	0,512836	0,000013	0,348400	0,000009	0,236404	0,000016
LG-07	0,704543	0,000011	0,056498	0,000006	0,513019	0,000011	0,348393	0,000009	0,236421	0,000014
LG-08	0,704523	0,000010	0,056498	0,000006	0,512671	0,000011	0,348396	0,000009	0,236407	0,000013
LG-10	0,703882	0,000008	0,056494	0,000005	0,512716	0,000013	0,348388	0,000008	0,236395	0,000013
LG-11	0,704249	0,000011	0,056490	0,000007	0,512685	0,000016	0,348401	0,000013	0,236508	0,000024
Replicate analyses of NBS-987	$^{87}\text{Sr}/^{86}\text{Sr} =$	0,710259	$\pm$	0,000025	(n = 7; 2σ)					
Replicate analyses of LaJolla	$^{143}\text{Nd}/^{144}\text{Nd} =$	0,511830	$\pm$	0,000020	(n = 4; 2σ)					
Replicate analyses of Jndi	$^{143}\text{Nd}/^{144}\text{Nd} =$	0,512087	$\pm$	0,000018	(n = 11; 2σ)					

Table S6. Highly siderophile element concentrations (ppb), Re-Os isotopic results and model ages of Los Gemelos mantle xenoliths.

Sample	Os (ppb)	Ir (ppb)	Ru (ppb)	Pt (ppb)	Pd (ppb)	Re (ppb)	¹⁸⁷ Re/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os _(i)	T _{RD} (Ga)	T _{MA} (Ga)	Model age average	γOs	Al ₂ O ₃	Yb _N	Sp Cr#	Ol Mg#
LG-01	0,0238	0,1388	0,4241	7,7066	0,0183	0,0093	0,03127	0,12325	0,12325	0,87	0,94	0,903155448	-4,9	1,13	0,06	0,20	0,903
LG-02	0,8000	1,3191	0,4809	6,1425	0,1374	0,0076	0,02555	0,12193	0,12193	1,05	1,11	1,080304253	-5,9	1,13	0,11	0,36	0,911
LG-03	0,5967	0,8991	0,3597	2,4683	0,0347	0,0081	0,02723	0,12181	0,12181	1,06	1,13	1,099649408	-6,0	1,06	0,11	0,28	0,908
LG-06	5,8536	1,2074	1,5097	3,4089	0,0307	0,0013	0,00437	0,12150	0,12150	1,11	1,12	1,112087617	-6,2	1,15		0,33	0,912
LG-07	0,2748	0,5264	0,6553	6,8998	0,4165	0,0173	0,05816	0,12502	0,12502	0,63	0,72	0,676244802	-3,5	1,80	0,37	0,28	0,906
LG-08	1,0633	1,0008	1,5080	4,4511	0,0480	0,0147	0,04942	0,12134	0,12134	1,13	1,27	1,200067209	-6,4	0,83	0,09	0,38	0,906

¹⁸⁷Os/¹⁸⁸Os are measured values while ¹⁸⁷Os/¹⁸⁸Os_(i) are initial ratios calculated to the basalt eruption age (0.45 Ma) and assuming a decay constant for ¹⁸⁷Re of 1,666 x 10⁻¹¹ (Smoliar et al., 1996).

The Re-depletion (T_{RD}) and standard (T_{MA}) model ages and γOs are calculated assuming ¹⁸⁷Os/¹⁸⁸Os and ¹⁸⁷Re/¹⁸⁸Os values for the primitive upper mantle of 0,1296 and 0,4353, respectively (Meisel et al., 2001).

PUM (Becker et al., 2006)						
	Os	Ir	Ru	Pt	Pd	Re
	3,9	3,5	7,0	7,6	7,1	0,35

PUM-normalized values						
	Os	Ir	Ru	Pt	Pd	Re
LG-01	0,0061	0,0397	0,0606	1,0140	0,0026	0,0266
LG-02	0,2051	0,3769	0,0687	0,8082	0,0193	0,0217
LG-03	0,1530	0,2569	0,0514	0,3248	0,0049	0,0231
LG-06	1,5009	0,3450	0,2157	0,4485	0,0043	0,0037
LG-07	0,0705	0,1504	0,0936	0,9079	0,0587	0,0494
LG-08	0,2726	0,2859	0,2154	0,5857	0,0068	0,0420

PUM - Meisel et al. (2001)

¹⁸⁷Os/¹⁸⁸Os = 0,1296

¹⁸⁷Re/¹⁸⁸Os = 0,4353

Table S7. Trace element concentrations of reference materials (NIST610 and BCR-2) used in LA-ICP-MS measurements.

	Reference Material					Reference material (RM)		Precision and accuracy	
	Limit of detection	NIST610 (run; n = 32)	1 σ	BCR-2 (run; n = 11)	1 σ	NIST610 (GeoRem 5211)	BCR-2	NIST610 (run/RM)	BCR-2 (run/RM)
Rb	0,0117	431,13	2,18	53,13	1,11	425,70	46,02	1,01	1,15
Ba	0,0091	424,13	1,95	653,43	3,78	452,00	683,90	0,94	0,96
W	0,0320	445,33	2,82	0,58	0,12	444,00	0,47	1,00	1,24
Th	0,0121	450,63	3,25	5,52	0,14	457,20	5,83	0,99	0,95
U	0,0133	457,12	2,62	1,78	0,08	461,50	1,68	0,99	1,06
Nb	0,0088	419,41	2,09	11,16	0,20	465,00	12,44	0,90	0,90
Ta	0,0075	376,62	2,30	0,60	0,03	446,00	0,79	0,84	0,76
La	0,0077	457,43	2,44	25,08	0,32	440,00	25,08	1,04	1,00
Ce	0,0082	447,81	1,69	52,74	0,66	453,00	53,12	0,99	0,99
Pb	0,0208	413,33	2,97	10,67	0,37	426,00	10,59	0,97	1,01
Pr	0,0060	429,83	2,47	6,28	0,11	448,00	6,83	0,96	0,92
Mo	0,0345	376,81	2,09	249,99	2,97	417,00	250,60	0,90	1,00
Sr	0,0083	497,42	2,46	327,37	3,19	515,50	337,40	0,96	0,97
Nd	0,0319	430,82	2,89	27,57	0,30	430,00	28,26	1,00	0,98
Sm	0,0318	450,52	2,49	6,43	0,17	453,00	6,55	0,99	0,98
Zr	0,0134	439,91	2,16	164,28	3,31	448,00	186,50	0,98	0,88
Hf	0,0319	417,74	3,73	4,39	0,30	435,00	4,97	0,96	0,88
Eu	0,0120	461,13	3,03	1,95	0,08	447,00	1,99	1,03	0,98
Ti	0,3296	434,04	3,30	13186	247	452,00	13575	0,96	0,97
Gd	0,0316	419,92	2,97	5,56	0,21	449,00	6,81	0,94	0,82
Tb	0,0065	442,81	2,59	0,93	0,05	437,00	1,08	1,01	0,86
Dy	0,0270	426,52	3,40	5,75	0,26	437,00	6,44	0,98	0,89
Li	0,1261	484,64	4,01	10,84	0,52	468,00	9,13	1,04	1,19
Y	0,0070	449,92	2,38	29,58	0,62	462,00	36,07	0,97	0,82
Ho	0,0073	449,42	2,87	1,15	0,06	449,00	1,31	1,00	0,88
Er	0,0219	426,02	2,97	3,07	0,15	455,00	3,67	0,94	0,84
Tm	0,0054	420,13	2,86	0,44	0,02	435,00	0,53	0,97	0,83
Yb	0,0302	461,51	3,01	3,18	0,23	450,00	3,39	1,03	0,94
Lu	0,0077	434,73	2,88	0,45	0,03	439,00	0,50	0,99	0,89

Table S8. Whole-rock major and trace element concentrations of reference materials (W-2a, JP-1A, JP-1B, AGV-2, BHVO-2) used in iCAP-Qc ICP-MS measurements.

Sample	W-2a n=3	Reference values	RSD %	JP-1 n=3	Reference values	RSD %	AGV-2 n=3	Reference values	RSD %
wt%									
SiO2	52,681	<i>52,680</i>	0,002	42,40	<i>42,38</i>	0,04	57,10	<i>59,30</i>	2,67
Al2O3	15,450	<i>15,450</i>	0,001	0,68	<i>0,66</i>	1,67	16,85	<i>16,91</i>	0,24
Fe2O3 (t)	10,830	<i>10,830</i>	0,003	8,43	<i>8,37</i>	0,50	6,80	<i>6,69</i>	1,14
MnO	0,167	<i>0,167</i>	0,004	0,12	<i>0,12</i>	1,95	0,10	<i>0,10</i>	1,58
MgO	6,370	<i>6,370</i>	0,003	44,60	<i>44,60</i>	0,00	1,71	<i>1,79</i>	3,37
CaO	10,863	<i>10,860</i>	0,018	0,77	<i>0,55</i>	23,22	5,31	<i>5,20</i>	1,41
Na2O	2,200	<i>2,200</i>	0,001	0,04	<i>0,02</i>	36,18	4,09	<i>4,19</i>	1,62
K2O	0,633	<i>0,626</i>	0,831	bdl			3,40	<i>2,88</i>	11,79
TiO2	1,060	<i>1,060</i>	0,002	0,00	<i>0,01</i>	57,53	1,01	<i>1,05</i>	2,79
P2O5	0,140	<i>0,140</i>	0,021	0,00	<i>0,00</i>	1,42	0,31	<i>0,48</i>	29,66

Reference values (red and cursive) correspond to compiled recommended values in GeoRem (<http://georem.mpch-mainz.gwdg.de>), except for compiled proposed values, which are in black and cursive.

n: number of times analyzed; bdl: below detection limit

Table S8. Continued.

Sample	W2-a	Reference values ¹	RSD %	AGV-2	Reference values ¹	RSD %	BHVO-2	BHVO-2	Reference values ²	RSD %	JP-1 A	JP-1 B	Reference values ³		RSD %
ng/g	n=3														
Sc	36074	36074	0	13230	12819	2	32581	32100	31830	1	7316	7306	7240		1
Ti	6355859	6354611	0	6253171	6170823	1	16955478	16453389	16480000	1	20267	20180	21300		3
V	261654	261597	0	115977	116159	0	322791	313807	318200	0	25024	25433	27600		5
Co	44526	44526	0	15814	15846	0	45438	44487	44890	0	115205	115349	116000		0
Ni	69994	69993	0	18221	17923	1	118488	115423	119800	1	2306750	2290350	2460000	***	3
Cr	92792	92791	0	16364	15607	3	297048	290882	287200	1	2898549	2925692	2807000	***	0
Cu	103005	103000	0	49527	50125	1	126985	123507	129300	2	4805	4493	4250	**	6
Zn	77000	77000	0	88297	89254	1	99033	97770	103900	3	41996	42009	42000		0
Ga	17424	17424	0	20100	20377	1	21077	20909	21370	1	535	537	467	***	1
Rb	19804	19803	0	67706	68270	1	9174	9096	9261	1	311	304	340	***	2
Sr	194829	194828	0	660298	659951	0	395627	393128	394100	0	568	551	570	*	2
Y	20114	20113	0	18418	18321	0	24478	24311	25910	3	94	105	100	*	6
Zr	87869	87866	0	229804	231199	0	169167	167800	171200	1	5201	5663	5920	***	5
Nb	7276	7275	0	14149	14004	1	18481	18304	18530	0	71	37	54	**	31
Cd	77	77	0	114	121	4	107	106	115	4	7	7	9	***	7
Sn	1950	1950	0	2100	2242	5	1910	1977	1776	4	60	49	50		11
Cs	888	888	0	1163	1168	0	101	96	97	1	39	39	35	***	2
Ba	169683	169680	0	1137447	1148663	1	131615	130967	130900	0	10169	10124	9800	*	2
La	10521	10521	0	37814	38412	1	15151	15109	15200	0	28	28	30	***	1
Ce	23217	23216	0	69303	69635	0	37441	37427	37760	0	61	61	60	***	0
Pr	3025	3025	0	8243	8268	0	5360	5364	5339	0	8	8	7	***	2
Nd	12911	12911	0	30381	30538	0	24296	24349	24780	1	32	30	33	***	3
Sm	3266	3266	0	5476	5505	0	6022	6034	6023	0	8	8	8	**	2
Eu	1094	1094	0	1528	1519	0	2058	2073	2043	1	2	2	2	**	5
Gd	3708	3708	0	4527	4533	0	6252	6226	6207	0	10	10	10	**	4
Tb	615	615	0	630	632	0	934	940	939	0	2	2	2	**	3
Dy	3808	3808	0	3439	3457	0	5232	5225	5280	0	13	14	15	**	6
Ho	803	803	0	676	679	0	997	1002	989	1	3	4	4	**	9
Er	2222	2222	0	1795	1807	0	2501	2508	2511	0	12	13	14	***	7
Tm	327	327	0	258	261	1	336	336	335	0	2	3	3	***	6
Yb	2058	2058	0	1626	1641	1	1965	1974	1994	1	20	22	21	*	5
Lu	301	301	0	245	246	0	272	272	275	1	4	5	5	*	8
Hf	2356	2356	0	5048	5116	1	4307	4343	4470	2	118	142	126	**	10
Ta	454	454	0	833	835	0	1149	1151	1154	0	4	5		***	17
Pb	7528	7528	0	12710	12702	0	1723	1561	1653	5	82	88	90	*	6
Th	2105	2104	0	5896	5984	1	1172	1181	1224	2	11	14	13	**	10
U	505	505	0	1868	1893	1	416	421	412	1	12	28	12	**	55

¹ Reference values correspond to compiled recommended values in GeoRem (<http://georem.mpch-mainz.gwdg.de>)² Reference values from Jochum et al. (2016)³ Reference values from Imai et al. (1995), *Makishima and Nakamura (1997); **Makishima and Nakura 2006, ***Babechuk et al. (2010)

Table S9. Relevant information about mantle xenoliths from Los Gemelos, such as: rock-types; Mg# for silicates, oxide (Cr# for spinel), and adakite melts; Ca/Al ratios for clinopyroxene; Ce/YbN ratios for clinopyroxene and phlogopite (also La/PrN for clinopyroxene); counting point; P-T estimates; and partial melting estimates.

Sample	Rock types	Olivine	Orthopyroxene	Clinopyroxene			Spinel		Phlogopite (euhedral)		Phlogopite (inclusion)	Adakite glass veins	Modal proportion												P _{2PX} (BK ₉₀)	F % (WR)	F % (Cpx)			
		Mg#	Mg#	Mg#	Ca/Al	(La/Pr) _N	(Ce/Yb) _N	Cr#	Mg#	Mg#	(Ce/Yb) _N	Mg#	Mg#	Ol	Opx	Cpx	Sp	Phl	Sum	Opx/Cpx	Ol/Cpx	T _{2PX} (BK ₉₀)	T _{2PX} (TA ₉₈)	T _{Ca-in-OPX} (BK ₉₀)	T _{CPX} (NT ₀₀)	T _{AVERAGE}	T _{SD}			
LG01	lherzolite	0,90	0,91	0,92	4,83	-	-	0,20	0,79	-	-	-	-	64,5	21,9	9,0	4,6	-	100,0	2,4	7,1	919	888	908	874	897	20	15,6	28	-
LG02	harzburgite	0,91	0,92	0,94	8,05	1,60	0,32	0,28	0,75	-	-	-	0,57	58,5	35,9	3,9	1,6	-	100,0	9,2	15,0	849	839	838	831	839	8	19,0	24	17
LG03	harzburgite	0,91	0,91	0,94	7,72	4,40	0,02	0,28	0,77	-	-	-	-	76,2	17,1	3,2	3,5	-	100,0	5,3	23,8	871	865	860	857	863	6	17,4	24	16
LG04	harzburgite	0,91	0,92	0,93	5,18	-	-	0,42	0,67	-	-	-	-	82,5	9,8	3,2	4,5	-	100,0	3,1	26,0	1020	977	804	999	950	99	22,4	-	-
LG05	harzburgite	0,91	0,91	0,94	7,91	-	-	0,29	0,77	-	-	-	-	79,6	13,8	2,1	4,5	-	100,0	6,6	37,9	874	868	861	858	865	7	17,8	22	-
LG06	harzburgite	0,91	0,92	0,94	8,50	1,50	0,27	0,33	0,74	-	-	-	0,62	60,2	34,9	3,3	1,6	-	100,0	10,5	18,1	780	772	821	758	783	27	19,8	-	17
LG07	harzburgite	0,91	0,91	0,93	8,07	-	-	0,28	0,74	-	-	-	-	71,4	22,9	3,3	2,4	-	100,0	6,9	21,4	803	800	861	784	812	34	17,6	-	-
LG08	harzburgite	0,91	0,91	0,94	7,86	3,24	10,11	0,38	0,73	0,92	15,32	0,92	0,57	82,2	10,5	2,5	3,9	0,9	100,0	4,2	32,9	923	897	773	902	874	68	21,6	26	16
LG09	harzburgite	0,91	0,91	0,94	8,16	6,89	0,05	0,29	0,78	-	-	-	-	79,6	14,0	2,7	3,7	-	100,0	5,2	29,6	877	869	856	866	867	9	18,5	-	17
LG10	lherzolite	0,91	0,91	0,93	5,01	-	-	0,21	0,80	-	-	-	-	67,8	20,0	8,5	3,7	-	100,0	2,4	8,0	927	896	902	889	903	17	15,8	28	-
LG11	harzburgite	0,91	0,91	0,94	7,97	4,04	0,02	0,28	0,77	-	-	-	-	73,7	18,1	4,6	3,6	-	100,0	3,9	16,0	902	892	862	891	887	17	18,7	28	17
LG12	harzburgite	0,91	0,91	0,94	7,84	2,58	8,79	0,38	0,72	0,92	22,08	0,92	0,56	84,0	10,7	1,5	3,5	0,3	100,0	7,0	55,3	823	798	820	792	808	15	21,6	12*	16
LG13	harzburgite	0,92	0,92	0,93	8,16	-	-	0,29	0,74	-	-	-	-	77,6	16,2	3,5	2,7	-	100,0	4,6	22,2	707	699	811	675	723	60	17,6	-	-
LG14	harzburgite	0,92	0,92	0,95	8,94	1,54	0,30	0,38	0,76	-	-	-	0,64 / 0,59	69,1	26,5	2,4	2,0	-	100,0	11,0	28,8	811	798	852	789	813	28	20,2	-	17
LG15	harzburgite	0,91	0,92	0,94	8,67	1,71	0,27	0,31	0,74	-	-	-	0,63 / 0,62	66,2	29,8	3,1	0,9	-	100,0	9,6	21,2	821	810	833	802	816	13	20,3	-	17
LG16	harzburgite	0,91	0,91	0,93	5,77	4,25	0,01	0,32	0,78	-	-	-	-	72,7	21,7	2,4	3,1	-	100,0	8,9	29,9	1011	987	860	1006	966	71	20,1	-	16

Mg# = Mg/(Mg+Fe)
Cr# = Cr/(Cr+Al)
Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene, Sp = spinel, Phl = phlogopite, WR = whole-rock
*We assume this estimate as unrealistic

CAPÍTULO 4

CONSIDERAÇÕES FINAIS

4 CONSIDERAÇÕES FINAIS

Este trabalho busca corroborar com o entendimento da complexa e longuíqua evolução do manto litosférico subcontinental (SCLM) abaixo do retroarco patagônico, através de análises de xenólitos mantélicos, aos quais apresentam espinélio em sua composição, oriundos do vulcão Los Gemelos, Meseta de Canquel, com idades-modelo Re (T_{RD}) no intervalo de 1,20 a 0,68 bilhões de anos (Ga) correspondentes ao Meso-Neoproterozoico. Os xenólitos analisados, podem ser classificados como harzburgitos ($n = 14$) e lherzolitos pobres em clinopiroxênio ($n = 2$), sendo caracterizados por altos graus fusão parcial (16–28%) e registram enriquecimento sobreposto (metassomatismo).

O empobrecimento é evidenciado pelos baixos teores de Al_2O_3 e CaO na rocha total (ambos $<1,5\%$ em peso), alto Mg# na olivina (>90) e espinélio rico em Cr ($Cr\# >0,20$), além de outros indicadores envolvendo correlações típicas entre o Mg# e elementos basálticos selecionados (por exemplo, CaO, Al_2O_3 , Na_2O e TiO_2) em rocha total e nos piroxênios. Além disso, os padrões de HSE (PGEs + Re) apoiam essa suposição, com base na composição empobrecida em elementos incompatíveis (Re e Pd). As idades modelo de Re-Os e as correlações positivas entre $^{187}Os/^{188}Os$ e o teor de Al_2O_3 na rocha total, sugerem que o evento de fusão parcial relacionado a formação e estabilização do SCLM sob Los Gemelos ocorreu durante o Mesoproterozóico (1,6–1,0 Ga).

O enriquecimento observado no SCLM sob o vulcão Los Gemelos é refletido no metassomatismo modal e críptico. Sendo o metassomatismo modal evidenciado pela cristalização de flogopita e pela ocorrência abundante de veios de vidro, bem desenvolvidos e interligados, com composição alcalina, contendo baixo e alto teores de SiO_2 . Quanto ao metassomatismo críptico, ele é evidenciado pelo enriquecimento seletivo de LREE (La-Ce) em relação aos outros LREEs no clinopiroxeno ($La/Pr_N = 1,50–6,89$) e padrão enriquecido em LREE na flogopita ($Ce/Yb_N = 15,32–22,08$), bem como pelo padrão *U-shaped* nas amostras de rocha total ($La/Sm_N >1$ e $Gd/Yb_N <1$).

Diagramas de discriminação selecionados, incluindo elementos maiores (ou seja, Mg# vs. elementos basálticos) e elementos traço (ou seja, La/Yb_N vs. Sr/Y e Ti/Eu ; e Sr vs. Ti/Eu) dos piroxênios, sugerem fortemente o papel de um *melt* silicático com composição adakítica como um candidato provável para agente metassomático responsável pelo enriquecimento (em extensões variáveis) do SCLM sob o vulcão Los Gemelos. É importante salientar que a intensidade do metassomatismo diminui do manto mais profundo (~ 70 km) em direção à seção

mais rasa (~50 km). Os melts em equilíbrio calculados com os clinopiroxênios revelam que o G1 (ao qual coexistem com flogopita) foi fortemente metassomatizado por um líquido adakítico, enquanto G2 e G3 foram ligeiramente afetados por esse evento metassomáticos. Os isótopos de Sr-Nd em rocha total, embora não conclusivos, sugerem a adição de componentes típicos relacionados à subducção. Portanto, é razoável concluir que a cristalização da flogopita e o metassomatismo críptico, registrados pelos xenólitos mantélicos, são produtos da fusão parcial da crosta oceânica jovem, e, portanto, quente, subduzida, que levou a geração de líquidos adakíticos como agentes metassomáticos na cunha do manto. Entre os prováveis eventos tectônicos capazes de gerar o metassomatismo adakítico, identificado nos xenólitos mantélicos de Los Gemelos, consideramos o desenvolvimento de uma janela de subducção durante o Paleoceno ao Eoceno como um candidato compatível.

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