UNIVERSIDADE FEDERAL DO RIO GRANDE DO SUL
CENTRO ESTADUAL DE PESQUISA EM SENSORIAMENTO REMOTO E METEOROLOGIA
PROGRAMA DE PÓS-GRADUAÇÃO EM SENSORIAMENTO REMOTO

IDENTIFICAÇÃO DA INFLUENCIA DO EL NIÑO – OSCILAÇÃO SUL E OSCILAÇÃO DECENAL DO PACÍFICO SOBRE AS GELEIRAS ANDINAS TROPICAIS USANDO SENSORIAMENTO REMOTO E PARÂMETROS CLIMÁTICOS

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Porto Alegre (RS), Janeiro de 2017

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Tese de Doutorado submetida ao Programa de Pós-Graduação em Sensoriamento Remoto da Universidade Federal do Rio Grande do Sul como requisito parcial para obtenção do título de Doutor em Sensoriamento Remoto.

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PORTO ALEGRE
2017

Veettil, Bijeesh Kozhikkodan
Identificação da influência do El Niño-Oscilação Sul e Oscilação Decenal do Pacífico sobre as geleiras Andinas Tropicais usando Sensoriamento Remoto e parâmetros climáticos. / Bijeesh Kozhikkodan Veettil. - Porto Alegre: IGEO/UFRGS, 2017.
[252 f.] il.

Tese (Doutorado).- Universidade Federal do Rio Grande do Sul. Programa de P· s-Graduação em Sensoriamento Remoto. Instituto de Geociências. Porto Alegre, RS - BR, 2017.

Orientador(es):Sergio Florêncio de Souza
Coorientador(es):Ulisses Franz Bremer

1. Andes tropicais 2. El Niño - Oscilação Sul (ENOS) 3. Oscilação Decenal do Pacífico (ODP) 4. Retração da geleira I. Título.

CDU 528.8

Catalogação na Publicação
Biblioteca Instituto de Geociências - UFRGS
Sônia Teresinha Duarte de Oliveira CRB 10/2310

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PORTO ALEGRE
2017

Dedico este trabalho à minha mãe
Gouri e ao meu pai Balan Nair

AGRADECIMENTOS

Ao meu orientador, ***Prof. Dr. Sergio Florêncio de Souza*** e ao meu coorientador ***Prof. Dr. Ulisses Franz Bremer***, meus agradecimentos pela ajuda, apoio, confiança, dedicação e paciência.

À ***Fundação de Amparo à Pesquisa do Estado do Rio Grande do Sul (FAPERGS)*** pela bolsa de doutorado concedida.

À ***Universidade Federal do Rio Grande do Sul (UFRGS)*** e ao ***Centro Estadual de Pesquisa em Sensoriamento Remoto e Meteorologia (CEPSRM)*** pelo ensino, pela qualidade e pela estrutura disponibilizada.

Ao ***Prof. Dr. Jefferson Cardia Simões*** e ***Centro Polar e Climático (CPC)*** pela estrutura disponibilizada.

À ***Dra. Shanshan Wang***, Key Laboratory of Arid Climate Change and Reducing Disaster of Gansu Province and Key Open Laboratory of Arid Climate Change and Disaster Reduction of CMA, Institute of Arid Meteorology CMA, Lanzhou, China e ao ***Prof. Dr. Éder Maier***, Universidade Federal do Rio Grande (FURG), Rio Grande pela ajuda com os dados climáticos e colaboração.

Ao ***Prof. Dr. Cláudio Wilson Mendes Jr*** pela participação na banca.

Ao ***Prof. Dr. Atilio Grondona*** pelas dicas sobre o curso e a tese, e pela amizade.

Aos meus queridos ***Nilceia*** e ***Marco Antonio*** pela paciência e pela compreensão.

À ***Geana*** pela sua ajuda desde inicio de inscrição de seleção de doutorado e pela amizade.

À ***Profa. Dra. Eliana Lima da Fonseca*** para o seu apoio no início do curso.

Aos professores, colegas, verdadeiros amigos e funcionários do CPC.

Aos professores, colegas, verdadeiros amigos e funcionários do CEPSRM.

“Somos o que nossos pensamentos fizeram de nós; portanto tome cuidado com o que você pensa. As palavras são secundárias. Os pensamentos vivem; eles viajam longe”

Swami Vivekananda (1863-1902)

RESUMO

Nas últimas décadas, particularmente desde a década de 1970, testemunhou-se um rápido recuo das geleiras em várias partes dos Andes tropicais. Uma tendência de aquecimento foi observada na região durante o mesmo período, com um hiato recente desde no início de 2010. No entanto, este hiato pode não ser o principal fator a influenciar as observações de aquecimento e recuo das geleiras em altitudes elevadas nos Andes tropicais. Com o surgimento de imagens de alta resolução espacial e espectral, e de modelos digitais de elevação (MDE) de alta resolução, agora é possível compreender as mudanças multitemporais das geleiras, o que era difícil de realizar utilizando as técnicas tradicionais e os dados de baixa resolução. Neste trabalho foram calculadas as variações da linha de neve das geleiras selecionadas ao longo dos Andes tropicais desde o início de 1980. A linha de neve máxima observada durante a estação seca (inverno austral) nos trópicos pode ser considerada como equivalente à linha de equilíbrio que separa a zona de acumulação da zona de ablação. A fim de reduzir o erro na estimativa da linha de neve foram consideradas somente as geleiras com declividades menores que 20º. Dependendo da região estudada e da presença de cobertura de nuvens, foram selecionadas imagens de várias fontes. As imagens da série Landsat (MSS, TM, ETM+ e OLI), EO1 OLI, ASTER e IRS LISS III foram usadas junto com MDE do ASTER GDEM-v2. Três bandas espectrais (TM5 - infravermelho médio, TM4 – infravermelho próximo e TM2 - verde) foram utilizadas para calcular a linha de neve durante a estação seca, aplicando limiares adequados para TM4 e TM2. Os conjuntos de dados meteorológicos de várias fontes também foram analisados para observar as mudanças na precipitação, na temperatura e na umidade que influenciam os parâmetros glaciológicos como: o balanço de massa e a linha de equilíbrio. Geleiras representativas nos trópicos internos e trópicos externos foram consideradas separadamente dentro de um novo quadro, que foi baseado na precipitação, umidade e condições de temperatura ao longo da América do Sul. Neste âmbito, os Andes tropicais são classificados em trópicos internos, trópicos externos úmidos do norte, trópicos externos úmidos do sul e os trópicos externos secos. O Vulcão Cotopaxi no Equador (trópicos internos), o Nevado Caullaraju-Pastoruri que é uma geleira na Cordilheira Branca no Peru (trópicos externos úmidos do norte), o Nevado Cololo na Cordilheira Apolobamba na Bolívia (trópicos externos úmidos do sul), o Nevado Coropuna na Cordilheira Ampato no Peru e o Nevado Sajama na Cordilheira Ocidental da Bolívia (trópicos externos secos) são as geleiras representativas de cada grupo consideradas neste estudo. As geleiras tropicais nos trópicos internos, especialmente as situadas perto da Zona de Convergência Intertropicais (ZCIT), são mais vulneráveis a aumentos na temperatura e menos sensíveis a variações na precipitação. Em contraste, as geleiras nos trópicos externos respondem à variabilidade de precipitação muito rapidamente em comparação com a variação de temperatura, particularmente quando se deslocam para as regiões subtropicais. A dependência do balanço de massa sobre as características de sublimação também aumenta a partir dos trópicos internos para os trópicos externos. As condições de aquecimento, com maior umidade, tendem a aumentar a perda de massa por causa do derretimento em vez da sublimação. A elevação da umidade nos trópicos externos pode alterar as geleiras dominadas pela sublimação (nos trópicos externos e subtrópicos) e para as geleiras dominadas por derretimento. Observa-se que as geleiras próximas da ZCIT (trópicos internos e trópicos externos úmidos do sul) estão recuando mais rapidamente como uma resposta ao aquecimento global, enquanto que as geleiras nos trópicos externos úmidos do norte e trópicos externos secos mostraram recuo relativamente mais lento. Possivelmente isso pode ser devido à ocorrência de fases frias do El Niño - Oscilação Sul (ENOS) conjuntamente com a Oscilação Decenal do Pacífico (ODP). As anomalias observadas nas variáveis meteorológicas seguem os padrões de ODP e as variações anuais de linha de neve seguem eventos de El Niño particularmente na fase ODP quente. No entanto, uma forte correlação entre as variações da linha de neve e dos fenômenos ENOS (e ODP) não está estabelecida. As geleiras do Equador mostram menos retração em resposta à tendência de aquecimento se comparadas às observações feitas por outros pesquisadores na Colômbia e na Venezuela, provavelmente devido à grande altitude das geleiras equatorianas. Em poucas palavras, as geleiras menores e em baixas altitudes nos trópicos internos e trópicos externos úmidos do sul estão desaparecendo mais rapidamente do que outras geleiras nos Andes tropicais. Também se observou neste estudo a existência de uma propriedade direcional no recuo das geleiras, o que não se observou em quaisquer outros estudos recentes. As geleiras nas cordilheiras leste do Peru e da Bolívia, que alimentam muitos rios nos lados leste das cordilheiras orientais, estão recuando do que aquelas geleiras situadas nas encostas ocidentais dos Andes tropicais.

Palavras chave: Andes tropicais, El Niño - Oscilação Sul (ENOS), Oscilação Decenal do Pacífico (ODP), retração da geleira, variações da linha de neve, Zona de Convergência Intertropical (ZCIT).

ABSTRACT

Recent decades, particularly since the late 1970s, witnessed a rapid retreat of glaciers in many parts of the tropical Andes. A warming trend is observed in this region during the same period, with a recent hiatus since the early 2010s. However, this hiatus is observed to have not influenced the retreat of high elevation glaciers in the tropical Andes. Due to the emergence of high spatial and spectral resolution images and high quality digital elevation models (DEM), it is now possible to understand the multi-temporal glacier changes compared with the techniques that existed a few decades before. We calculated the snowline variations of selected glaciers along the tropical Andes since the early 1980s. The maximum snowline observed during the dry season (austral winter) in the tropics can be considered as nearly equivalent to the equilibrium line that separates the accumulation zone from the ablation zone. In order to reduce the error in the estimated snowline, glaciers with slopes < 20^{o} only were considered in this research. Depending on the study region and the presence of cloud cover, images from multiple sources were selected. Landsat series (MSS, TM, ETM+, and OLI), EO1 OLI, ASTER, and IRS LISS III images were used along with digital elevation models (DEM) from ASTER GDEM-v2. Three wavebands (TM5 - Middle Infrared, TM4 - Near Infrared, and TM2 - Green) were used to calculate the dry season snowline, after applying suitable threshold values to TM4 and TM2. Meteorological datasets from multiple sources were also analysed to observe the changes in precipitation, temperature, and humidity that influence key glaciological parameters such as the mass balance and the equilibrium line. Representative glaciers in the inner and the outer tropical Andes were considered separately within a new framework, which is based on the precipitation, humidity, and temperature conditions along the South America. In this framework, tropical Andes are classified in to inner tropics, northern wet outer tropics, southern wet outer tropics, and dry outer tropics. Cotopaxi ice-covered volcano, Ecuador (inner tropics), Nevado Caullaraju-Pastoruri Glacier, Cordillera Blanca, Peru (northern wet outer tropics), Nevado Cololo, Cordillera Apolobamba, Bolivia (southern wet outer tropics), and Nevado Coropuna, Cordillera Ampato Peru and Nevado Sajama, Cordillera Occidental, Bolivia (dry outer tropics) are the representative glaciers in each group considered in this study. Inner tropical glaciers, particularly those situated near the January Intertropical Convergence Zone (ITCZ), are more vulnerable to increases in temperature and these glaciers are less sensitive to variations in precipitation. In contrast, outer tropical glaciers respond to precipitation variability very rapidly in comparison with the temperature variability, particularly when moving towards the subtropics. Mass balance dependency on sublimation characteristics also increases from the inner tropics to the outer tropics. Warming conditions with higher humidity tends to enhance mass loss due to melting rather than sublimation. Increased humidity observed in the outer tropics may change the sublimation dominated glaciers in the outer tropics and subtropics to melting dominated ones in the future. It is observed that the glaciers above and near the January ITCZ (inner tropics and southern wet outer tropics) are retreating faster as a response to global warming, whereas the glaciers in the northern wet outer tropics and dry outer tropics show relatively slower retreat. This can be possibly due to the occurrence of cold phases of El Niño-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) together. The observed anomalies in the meteorological variables slightly follow PDO patterns and the variations in annual snowlines follows El Niño events, particularly when in phase with warm PDO. However, a strong correlation between snowline variations and ENSO (and PDO) is not established. Mountain glaciers in Ecuador show less retreat in response to the warming trend compared with observations done by other researchers in Colombia and Venezuela, probably due to very high altitude of the Ecuadorean glaciers. In a nutshell, smaller glaciers at lower altitudes in the inner tropics and the southern wet outer tropics are disappearing faster than other glaciers in the tropical Andes. Another observation made in this study is the directional property of glacier retreat, which was not covered in any other recent studies. Those glaciers on the eastern cordilleras of Peru and Bolivia, which feed many rivers on the eastern sides of the eastern cordilleras, are retreating faster than those glaciers situated on the western sides.

Keywords: Tropical Andes, El Niño-Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), glacier retreat, snowline variations, Intertropical Convergence Zone (ITCZ).

APRESENTAÇÃO DA ESTRUTURA DA TESE

Esta Tese de Doutorado, intitulada "**IDENTIFICAÇÃO DA INFLUÊNCIA DO EL NIÑO – OSCILAÇÃO SUL E OSCILAÇÃO DECENAL DO PACÍFICO SOBRE AS GELEIRAS ANDINAS TROPICAIS USANDO SENSORIAMENTO REMOTO E PARÂMETROS CLIMÁTICOS"**, foi desenvolvida entre Março de 2013 e Março de 2016 no Centro Estadual de Pesquisas em Sensoriamento Remoto e Meteorologia (CEPSRM) da Universidade Federal do Rio Grande do Sul (UFRGS) em colaboração com o Centro Polar e Climático (CPC) da UFRGS.

A Tese é composta das seguintes partes:

Capítulo I: Aspectos introdutórios
Trata sobre os aspectos introdutórios e objetivos

Capítulo II: Revisão bibliográfica na forma de artigo
Artigo submetido intitulado "Glacier monitoring and glacier-climate interactions in the tropical Andes: a review", autores: BIJEESH KOZHIKKODAN VEETTIL, Shanshan Wang, Ulisses Franz Bremer, Sergio Florêncio de Souza, Jefferson Cardia Simões. Submetido para Journal of South American Eath Sciences.

Capítulo III: Metodologia

Capítulos IV, V, VI, VII e VIII: Resultados na forma de artigos científicos publicados
Capítulo IV. Artigo publicado intitulado "Combined influence of PDO and ENSO on northern Andean glaciers: a case study on the Cotopaxi ice-covered volcano, Ecuador", autores: BIJEESH KOZHIKKODAN VEETTIL, Éder Leandro Bayer Maier, Ulisses Franz Bremer, Sergio Florêncio de Souza. Publicado na Climate Dynamics 43, 3439-3448, 2014.

Capítulo V. Artigo publicado intitulado "Recent trends in annual snowline variations in the northern wet outer tropics: case studies from southern Cordillera Blanca, Peru", autores: BIJEESH KOZHIKKODAN VEETTIL, Shanshan Wang, Ulisses Franz Bremer, Sergio Florêncio de Souza, Jefferson Cardia Simões. Publicado na Theoretical and Applied Climatology, 2016.

Capítulo VI. Artigo publicado intitulado "Influence of ENSO and PDO in mountain glaciers in the outer tropics: case studied in Bolivia", autores: BIJEESH KOZHIKKODAN VEETTIL, Ulisses Franz Bremer, Sergio Florêncio de Souza, Éder Leandro Bayer Maier, Jefferson Cardia Simões. Publicado na Theoretical and Applied Climatology 125: 757-768, 2016.

Capítulo VII. Artigo publicado intitulado "Variations in annual snowline and area of na ice-covered stratovolcano in the Cordillera Ampato, Peru, using remote sensing data (1986-2014)", autores: BIJEESH KOZHIKKODAN VEETTIL, Ulisses Franz Bremer, Sergio Florêncio de Souza, Éder Leandro Bayer Maier, Jefferson Cardia Simões. Publicado na Geocarto International 31: 544-556, 2016.

Capítulo VIII. Artigo publicado intitulado "Un análisis comparativo del retroceso glaciar en los Andes tropicales usando teledetección", autores: BIJEESH KOZHIKKODAN VEETTIL, Sebastián Felipe Ruiz Pereira, Shanshan Wang, Pedro Teixeira Valente, Atilio Efrain Bica Grondona, Adriana Coromoto Becerra Rondón, Isabel Cristiane Rekowsky, Sergio Florêncio de Souza, Nilceia Bianchini, Ulisses Franz Bremer, Jefferson Cardia Simões. Publicado na Investigaciones Geográficas 51: 3-36, 2016.

Capítulo IX: Conclusões e considerações finais
Apresenta as conclusões e recomendações de futuras pesquisas.

SUMÁRIO

LISTA DE ABREVIATURAS E SIGLAS

AC: Altura de Congelamento

ADE: Aquecimento Dependente da elevação (Elevation Dependent Warming)

ALN: Altitude da linha de neve

a.n.m: acima do nível do mar

ASTER: Advanced Spaceborne Thermal Emission and Reflection Radiometer

CPC: Climate Prediction Centre (Centro de Previsão Climática)

CRU: Climate Research Unit (Unidade de Pesquisa Climática)

ENOS: El Nino – Oscilação Sul

ETM+: Enhanced Thematic Mapper+

GDEM: Global Digital Elevation Model

IOS: Índice de Oscilação Sul

ION: Índice Oceânico Niño

IPCC: Intergovermental Panel on Climate Change (Painel Intergovernamental sobre Mudanças Climáticas)

JISAO: Joint Institute for the Study of the Atmosphere and Ocean

LE: linha de equilíbrio

LISS 3: Linear Imaging Self Scanning Sensor 3

MDE: modelo digital de elevação

MSS: Multispectral Scanner

NDSI: Normalized Difference Snow Index (a diferença normalizada de índice de neve)

NOAA: National Oceanic and Atmospheric Administration

OA: Oscilação Antártica

ODP: Oscilação Decenal do Pacifico

OLI: Operational Land Imager

PEG: Pequena Era Glacial

PMM: precipitação média mensal

SCOR - WG 55: Scientific Committee on Ocean Research - Working Group 55

SENAMHI: Serviço Nacional de Meteorologia e Hidrologia

SRTM: Shuttle Radar Topographic Mission

TM: Thematic Mapper

TSM: temperatura da superfície do mar

USGS: United States Geological Survey

ZVC: Zona Vulcânica Central

CAPITULO I

Aspectos Introdutórios: Apresentação e Objetivos

1. APRESENTAÇÃO

O Painel Intergovernamental sobre Mudanças Climáticas (IPCC) reconheceu a função das geleiras de montanhas como o principal indicador de mudanças climáticas (Lemke *et al*., 2007). As geleiras tropicais em regiões montanhosas são extremamente sensíveis à mudança climática (Hastenrath, 1994) e as condições climáticas nessas regiões favorecem esta ablação durante todo o ano para o término das geleiras (Francou *et al*., 2004). Bradley *et al*. (2006) projetou um aumento de temperatura em torno de 4ºC em altitudes superiores a 4000 m acima do nível do mar (a.n.m) no século XXI. A presente pesquisa concentra-se na influência das mudanças climáticas recentes que ocorreram nos Andes tropicais (**figura 1**) usando sensoriamento remoto, para entender as diferenças na resposta das geleiras em locais diferentes nos Andes tropicais. As subseções seguintes descrevem a importância regional e global de geleiras andinas tropicais, a correlação entre as mudanças climáticas e tais geleiras, as recentes variações das mesmas nos países andinos tropicais (Equador, Peru e a Bolívia).

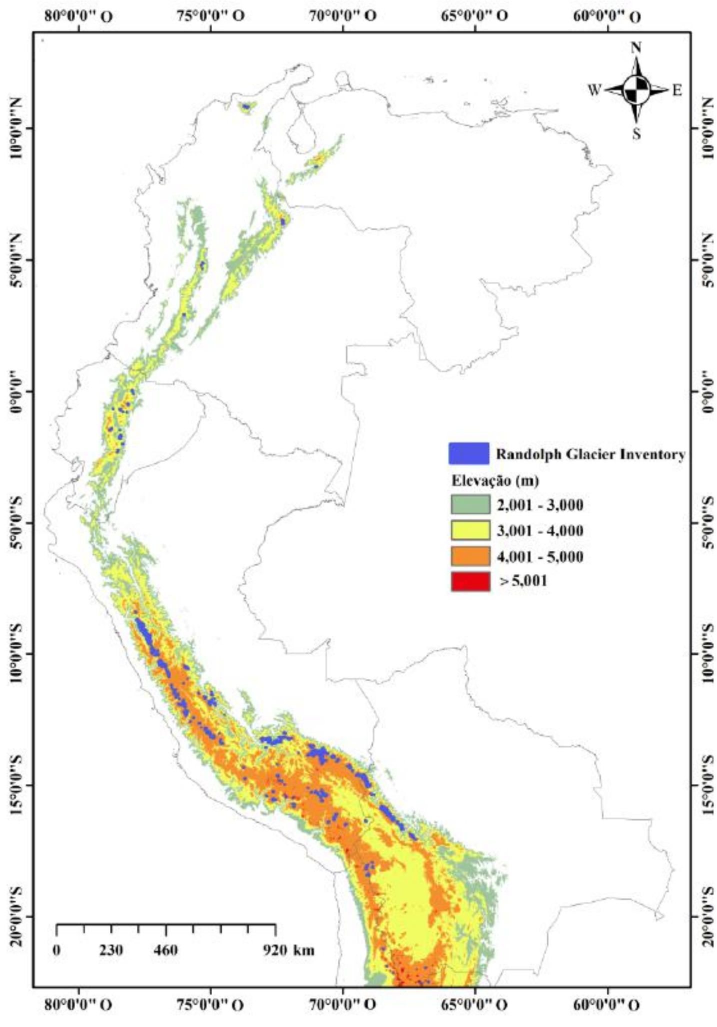

Figura 1—Os Andes tropicais.

1.1. Geleiras andinas tropicais

Mais de 80% do fornecimento de água doce para a população nas regiões áridas e semi-áridas das áreas tropicais e subtropicais têm sua origem nas regiões de montanha, que inicialmente estão classificadas como gelo em geleiras (Messerli, 2001; Vuille *et al*., 2008a) e cerca de 99,7% de todas as geleiras tropicais estão situadas na América do Sul, que recobre cerca de 2.758 km^2 (Kaser, 1999). Durante a estação seca, os países andinos, especialmente as áreas mais populosas na Bolívia e no Peru, dependem de água doce originarias das bacias glaciais. As geleiras tropicais, portanto, funcionam como amortecedor crítico contra a precipitação altamente sazonal e fornecem água para uso doméstico, agrícola ou industrial, durante a ausência/diminuição da chuva (Vuille *et al*., 2008a). Essas geleiras, também funcionam como um dos indicadores mais visíveis para mudanças climáticas devido ao rápido tempo de resposta, a sensibilidade às variações climáticas e a clara percepção na sua reação (perda ou ganho de balanço de massa) (Vuille *et al*., 2008a, 2008b). A zona tropical pode ser dividida em duas subzonas com base nas características de precipitação e localização geográfica (Rabatel *et al*., 2013). Na zona tropical interior (Colômbia e Equador), ao longo do ano, há precipitação contínua com ablação e acumulação que ocorrem simultaneamente. Na zona tropical exterior (Peru e Bolívia), durante a estação seca que ocorre de maio a setembro com as condições subtropicais e uma temporada molhada de outubro a março, quando as condições tropicais prevalecem (Rabatel *et al*., 2013) com notável acúmulo somente durante a estação chuvosa (Jomelli *et al*., 2009).

1.2. As mudanças climáticas e as geleiras andinas tropicais

Várias observações sobre as medidas das geleiras do Equador, Bolívia, Peru, Chile e Colômbia demonstraram que tem ocorrido uma contração rápida e descontinua das geleiras andinas tropicais desde a Pequena Idade do Gelo (PIG) (Vuille *et al*., 2008a). Essas geleiras estão constantemente próximas das condições de fusão particularmente sensíveis às mudanças climáticas devido à localização. O clima andino tropical também foi alterado ao longo das últimas 5 ou 6 décadas, em paralelo com o recuo das geleiras e um aumento de 0,1^{o}C/década foi identificado na temperatura atmosférica na região andina, bem como um ligeiro aumento na precipitação na segunda metade do século XX (Vuille *et al*., 2008b). Verifica-se que a circulação atmosférica tropical também aumentou recentemente e vários modelos de previsão climática indicaram um aquecimento contínuo da troposfera tropical ao longo do século XXI, com um aumento de temperatura reforçada nas altitudes mais elevadas (Vuille e Bradley, 2000; Vuille *et al*., 2003; Bradley *et al*., 2004, 2006). Além disso, os modelos também sugerem que a precipitação da estação seca pode aumentar efetivamente o ciclo hidrológico sazonal nos Andes tropicais (Vuille *et al*., 2008a). A reconstrução climática usando geleiras pode basear-se em depósitos de moraina glacial ou registros de testemunhas de gelo, enquanto outros métodos incluem a datação por radiocarbono das morainas glaciais (Clapperton, 1981; Rodbell, 1992; Rabatel *et al*., 2006) e liquenometria (Rodbell, 1992). O número de estudos para reconstruir as flutuações glaciais dos depósitos de moraina glacial nos Andes é muito limitado (Jomelli et al., 2009). O comprimento das geleiras, área de superfície e linha de equilíbrio (LE) são os parâmetros glaciológicos dominantes utilizados para previsões climáticas (Jomelli *et al*., 2009). A maioria das geleiras andinas foi monitorada com base em LE, sendo a definição tradicional de LE referente à altitude onde bn = 0, com bn sendo a ablação líquida ao final da temporada de ablação (verão) (Benn *et al*., 2005). No caso das geleiras tropicais, onde ocorre a ablação durante todo o ano, essa definição não é adequada (Benn *et al*., 2005). Greene *et al*. (2002) expressam LE em função do congelamento, altura e precipitação média anual como demonstrado abaixo:

$$LE = 537 + 1{,}01AC - 0{,}51P \qquad \text{Eq (1)}$$

Onde AC é a altura de congelamento (em metros, acima do nível do mar) e P é a precipitação média anual (em milímetros).

Polissar *et al*. (2006) calcularam um modelo de reconstrução de geleira onde ocorreram incertezas na estimativa da temperatura com base em LE para a Venezuela. Rabatel *et al*. (2006) reconstruíram a retração da geleira em Cerro Charquini nos Andes centrais na Bolívia entre a PIG e as variações nas médias LE em 1997. Apesar de não ser completa, a interpretação das variações na extensão da geleira ou LE, juntamente com dados de temperatura e precipitação são mais usados atualmente e desde as últimas décadas (Benn *et al*., 2005; Jomelli *et al*., 2009).

1.3. Variações das geleiras nos Andes tropicais

Nas subseções seguintes são descritos, brevemente, as variações glaciais em três países andinos: Equador, Bolívia e Peru.

1.3.1. Equador

As geleiras no Equador estão situadas nas duas cordilheiras - Cordilheira Ocidental (Ex. Iliniza e Chimborazo) e Cordilheira Oriental (Ex. Cotopaxi, Antisana e Cayambe). As geleiras equatorianas pertencem à região dos trópicos internos onde há a precipitação continua com a ablação e acumulação ocorre simultaneamente ao longo do ano. O Cotopaxi é o vulcão ativo mais alto do mundo (5.911 m de altitude.). Nesta região localiza-se 4% do gelo tropical do mundo, se encontra confinado aos estratovulcões isolados cobertos de gelo e compõem uma das principais fontes de água doce para a capital Quito. No futuro, a população equatoriana dos Andes estará vulnerável às mais imediatas e intensas mudanças climáticas que estão em curso, pois de acordo com La Frenierre (2012), a localização geográfica de elevada altitude e baixa latitude coloca as geleiras dentro de uma zona que tende a ser a primeira a sofrer os impactos das mudanças climáticas. Os registros históricos mostraram que houve uma extensa glaciação entre os séculos XV e XVIII, seguida por várias retrações que foram interrompidas devido aos avanços glaciais por volta de 1800, 1850 e 1870 (Jomelli *et al*., 2009). A LE subiu até 250 m entre o século XVIII e o presente (Vuille *et al*., 2008a). No século XX, foram encontradas geleiras que sofreram uma perda de área até 75% no Equador. A geleira Antizana 15, por exemplo, sofreu um grande recuo durante 1956-1998 (Francou *et al*., 2000). Jordan *et al.* (2005) calcularam que o Cotopaxi havia perdido 30% de sua área de superfície da geleira entre 1976 e 1997, com base no estudo sobre geleiras selecionadas. Perto do Cotopaxi, o vulcão Antizana perdeu até 33% de sua área glacial entre 1979 e 2007 (Rabatel *et al*., 2013). O vulcão Chimborazo, também reduziu em 57%, de 27,7 km^2 para 11,8 km^2 de área de superfície glacial durante o período de 1962-1997 (Cáceres, 2010). No Equador, as geleiras demonstram um pequeno avanço durante os períodos de La Niña, entre 1999 e 2001 (Francou *et al*., 2004).

1.3.2. Peru

A cobertura da terra no Peru pode ser dividida em três regiões geográficas: a costa do Pacífico, a Cordilheira dos Andes e a Floresta Amazônica (Chevallier *et al*., 2011). O derretimento das geleiras a partir dessas montanhas de grande altitude fornece água doce durante a estação seca. Cerca de 70% de todas as geleiras tropicais estão situadas nos Andes peruanos (Vuille *et al*., 2008a). Inúmeros documentos estão disponíveis pelo fato de que as geleiras menores na Cordilheira Branca no Peru estão desaparecendo nestes últimos anos. Os registros dos gelos do Peru mostraram a correlação direta entre ENOS e geleiras nos Andes peruanos (Henderson *et al*., 1999; Thompson *et al*., 2000; Herreros *et al*., 2009). Um dos registros de testemunhos de gelo mais documentado é o aquecimento e recuo de calota de gelo Quelccaya no Peru (Thompson *et al*., 2000). Estudos baseados em sensoriamento remoto previram que muitas geleiras no Peru, como Yanamarey Glaciar, vão desaparecer dentro de uma década (Huh *et al*., 2012). Foram encontrados avanços glaciais, que podem ter ocorrido em 1330 +/- 29 (Solomina *et al*., 2007), embora sobreposta pela PIG (Vuille *et al*., 2008a). A extensão máxima encontrada no Peru foi entre os anos de 1630 e 1680 devido à PIG com pelo menos três avanços durante os séculos XVII e XVIII (Vuille *et al*., 2008a). O clima do Peru é altamente influenciado pelas montanhas andinas. As alterações na LE denotam mudanças imediatas no balanço de massa da geleira, em escala interanual e sua mudança contínua pode ser usada para estimar a tendência climática na região. Recentemente inúmeros documentos foram disponibilizados pelo fato de que as geleiras menores na Cordilheira Branca, no Peru, estão desaparecendo (Vuille *et al*., 2008a; Jomelli *et al*., 2009; Chevallier *et al*., 2011; Sagredo e Lowell, 2012; Rabatel *et al*., 2013). Registros de testemunhos de gelo do Peru mostraram a correlação direta entre El Niño - Oscilação Sul (ENOS) e alterações de balanço de massa das geleiras nos Andes peruanos (Henderson *et al*., 1999; Thompson *et al*., 1984; Herreros *et al*., 2009). Um dos registros de testemunhos de gelo mais documentado é o que apresenta o recuo da calota de gelo Quelccaya, no Peru (Thompson *et al*., 1984). Entre 1990 e 2009, a redução anual encontrada na Cordilheira Branca no Peru foi de 0,81% (Baraer *et al*., 2012). A **figura 2** mostra alterações na extensão e área da superfície de 10 geleiras andinas tropicais no Equador, Peru e Bolívia.

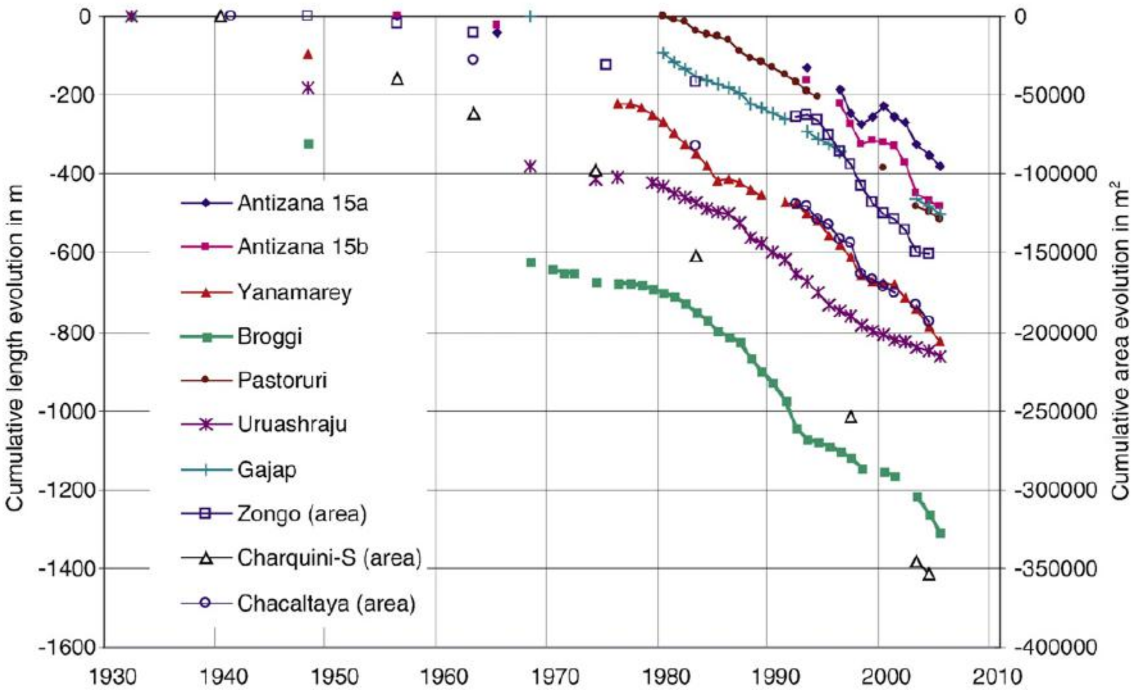

Figura 2—Alterações na extensão e área de superfície de 10 geleiras andinas tropicais do Equador (15 Antizana e 15b), Peru (Yanamarey, Broggi, Pastoruri, Uruashraju, Gajap) e Bolívia (Zongo, Charquini, Chacaltaya) entre 1930 e 2005. Fonte: Vuille *et al*. (2008a)

1.3.3. Bolívia

A Bolívia é considerada um país tropical, suas principais divisões de altitude consistem em terras baixas (< 800 m de altitude), encostas andinas (800-3200 m de altitude) e planaltos ou Altiplano (> 3.200-6.500 m a.n.m). As geleiras bolivianas estão situadas em duas cadeias de montanhas: a Cordilheira Ocidental ao longo da fronteira oeste com o Chile e as Cordilheiras Apolobamba, Real, Três Cruzes e Nevado de Santa Vera Cruz, no leste (Vuille *et al*., 2008a). As geleiras da região ocidental se restringem ao Nevado Sajama e aos vulcões na fronteira do Chile. A maioria das geleiras é composta por camadas de gelo, geleiras de vale e geleiras de montanha, e estão situadas na Cordilheira Oriental na Bolívia (Vuille *et al*., 2008a). O sul da Bolívia atualmente é desprovido de geleiras devido à precipitação baixa (Messerli *et al*., 1993). Os avanços máximos glaciais encontrados ocorreram durante a segunda metade do século XVII (Rabatel *et al*., 2006) e a grande retração ocorreu durante o século XX, especialmente após a década de 1940. Algumas das geleiras, como a Charquini, perderam da área durante a PIG até 78% e a LE subiu 160 m (Rabatel *et al*., 2006), embora nas maiores geleiras, como Zongo, tenha mostrado perda relativamente menor, provavelmente devido a precipitação reabastecida (Vuille *et al*., 2008a). Um dos desaparecimentos mais notáveis das geleiras ocorreu na Chacaltaya, que perdeu 62% de sua massa entre 1940 e 1983 (Francou *et al.*, 2000). A Chacaltaya pode ser considerada como um representante de muitas geleiras de tamanho semelhante na Cordilheira Real (Francou *et al.*, 2000). A variabilidade interanual e as tendências de longo prazo da maior geleira (Zongo) e da menor geleira (Chacaltaya), foram muito semelhantes (Vuille *et al.*, 2008a) (**figura 3**).

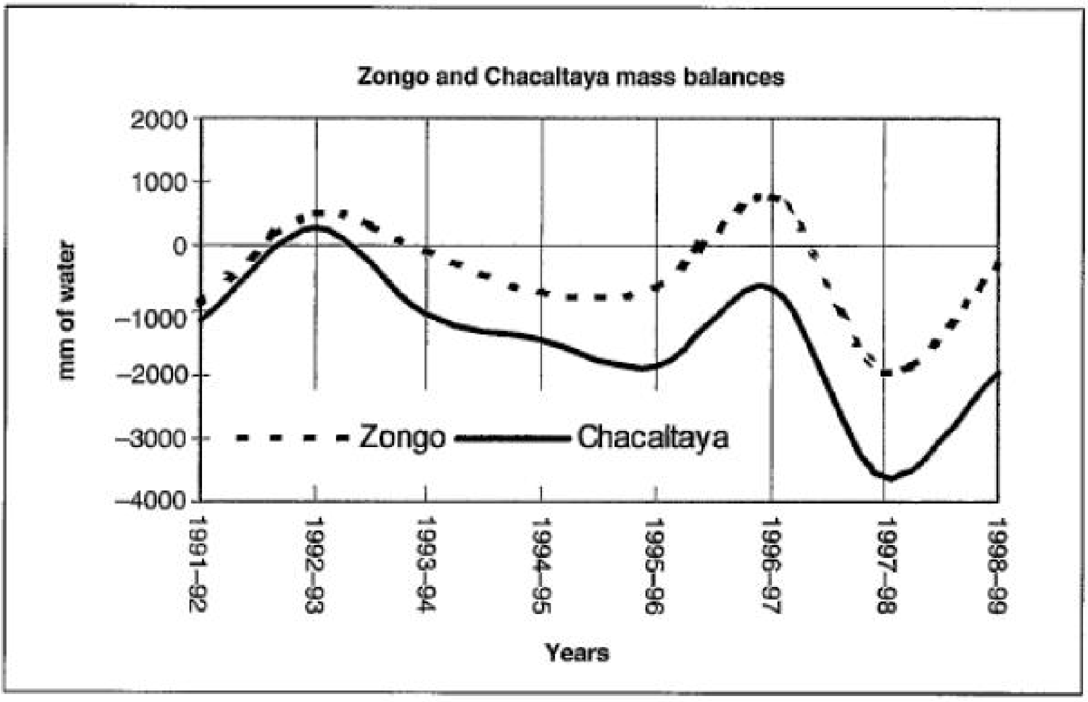

Figura 3—Evolução do balanço de massa das geleiras Zongo e Chacaltaya, em agosto de 1999. A perda acumulada totalizou 11.500 mm em Chacaltaya e 3.350 milímetros em Zongo Fonte: Francou *et al*. (2000)

1.4. El Niño - Oscilação Sul e as variações das geleiras

O termo "El Niño" foi aplicado originalmente a uma fraca corrente oceânica quente anual que corre para o sul ao longo da costa do Peru e do Equador durante o período do Natal e só posteriormente tornou-se associado com os grandes aquecimentos que vem ocorrendo há poucos anos e mudam a ecologia local e regional (Trenberth, 1997). O grupo de trabalho do Scientific Committee on Oceanic Research (SCOR WG 55), definiu o El Niño como o aparecimento de água anormalmente quente ao longo da costa do Equador e Peru até o sul de Lima (12°S). Isto significa uma anomalia de temperatura da superfície do mar (TSM) normalizada superior ao desvio-padrão em pelo menos quatro (4) meses consecutivos. Esta anomalia de TSM normalizada deve ocorrer em pelo menos em três (3) das cinco (5) estações costeiras peruanas. Nos prazos interanuais, uma fração significativa da variabilidade da precipitação está relacionada ao fenômeno ENOS (Rabatel *et al*., 2013; Vuille e Bradley, 2000), que é uma das fontes importantes de variabilidade climática interanual na Terra (Christie *et al*., 2009). Os anos de El Niño são normalmente associados às condições quentes e secas, enquanto os anos de La Niña estão associados com condições de frio e umidade no Altiplano, apesar das características climáticas do El Niño/La Niña não serem uniformes entre os Andes tropicais (Rabatel *et al*., 2013). Os Andes tropicais (entre 10ºN e 16ºS) são o conjunto de locais adequados para o estudo da influência das alterações climáticas nas geleiras tropicais. Estas geleiras estão sujeitas a maiores variações de temperatura diárias do que variações de temperaturas anuais. Sabe-se há muito tempo que os anos de ocorrência de El Niño foram seguidos pela diminuição de precipitação no norte da América do Sul e isto se deve à inibição do transporte de umidade da bacia Amazônica recorrente do forte vento de oeste durante El Niño (Vuille, 2013). Nota-se também que as vazões dos rios durante os períodos de El Niño intensos foram maiores em direção ao norte do Peru em comparação com os rios do sul (Casimiro *et al*., 2013). Vuille *et al*. (2000) propuseram que a variabilidade da precipitação em direção ao leste dos Andes equatoriais é mais relacionada com as anomalias de circulação tropical do Atlântico do que as do Pacífico. Outro ponto sobre a variabilidade de precipitação nos Andes é que esta aumenta com a altitude, como pode ser observado nas grandes altitudes dos Andes equatorianos (Garreaud *et al*., 2009). A precipitação é uma das variáveis que determina o crescimento/recuo das geleiras, particularmente na região tropical e, portanto é importante entender como a precipitação varia na região dos Andes. Nota-se também que a precipitação perto das geleiras próxima a Bacia Amazônica é maior do que a precipitação perto do Pacífico. Constatou-se que as anomalias de temperatura e precipitação associadas ao ENOS estão enfraquecendo do norte para o sul (Garreaud, 2009), provavelmente devido à maior influência dos padrões de circulação sobre o Atlântico tropical ou a Bacia Amazônica. Sabe-se também que o El Niño - Oscilação Sul e outros fenômenos de escala mundial, como a Oscilação Decenal do Pacífico (ODP) influenciam o clima andino de forma diferente ao longo de sua ocorrência (Garreaud, 2009).

1.5. Oscilação Decenal do Pacifico (ODP)

A Oscilação Decenal do Pacífico (ODP) é um índice de clima com base nas variações do TSM do Pacífico Norte (Mantua *et al*., 1997) com água morna (índice positivo) e regimes de frio (índice negativo). A ODP tem sido descrita por alguns pesquisadores como sendo um padrão de longo prazo de El Niño de variabilidade climática do Pacífico (Mantua e Hare, 2002). Os regimes quentes e frios podem persistir por várias décadas. Desde 1998, estes ciclos decenais foram rompidos. Durante a ODP: (a) Ocorre uma fase positiva ou quente, caracterizada por anomalias positivas ao longo da Costa Oeste da América do Norte e por anomalias negativas no resto da bacia do Pacífico Norte, entre aproximadamente 20ºN e 90ºN na direção do polo (neste caso a forma de cálculo do índice conduz a valores positivos); (b) uma fase negativa ou fria quando o aquecimento/resfriamento nas duas áreas se inverte (neste caso o índice tem resultado negativo) (**figura 4**).

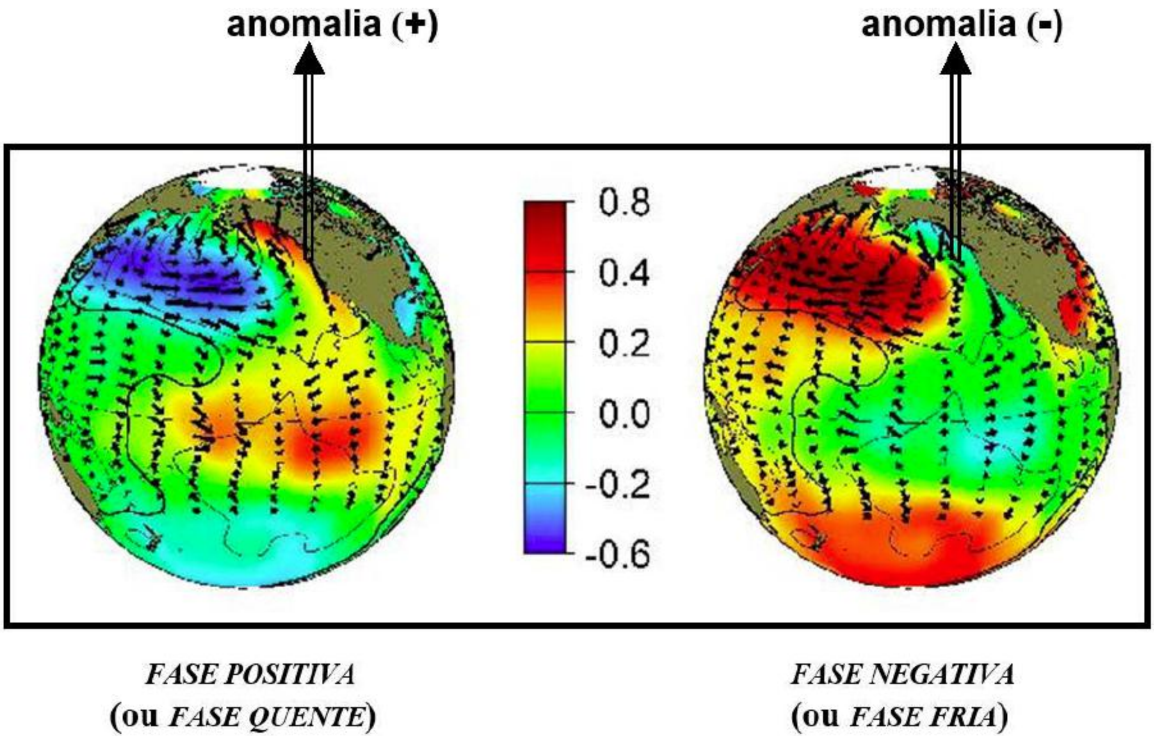

Figura 4—As fases da ODP. Fonte: Mantua *et al*. (1997)

Nota-se, que a ODP comporta-se como uma "gangorra" entre a TSM ao longo da costa oeste da América do Norte, com relação ao restante da bacia do Pacífico Norte. Da mesma maneira como o Índice de Oscilação Sul (IOS) constitui também uma "gangorra", no Pacífico Sul, no que se refere às variações da pressão atmosférica de superfície entre dois pontos ao leste e ao oeste da bacia do Pacífico Sul (por exemplo, Tahiti, mais próximo à costa da América do Sul, e Darwin, na Austrália). A ODP pode modular a relação interanual entre o ENOS e clima global e também pode causar variabilidade climática interdecadal no Pacífico tropical (Mantua *et al*., 1997). Duas características principais distinguem a ODP de ENOS.

Em primeiro lugar, eventos típicos de ODP têm mostrado notável persistência em relação aos eventos ENOS - neste século, as grandes eras de ODP persistiram de 20 a 30 anos (Mantua *et al*., 1997; Minobe 1997). Em segundo lugar, as marcas climáticas da ODP são mais visíveis no Pacífico Norte, enquanto assinaturas secundárias existem nos trópicos - o oposto é verdadeiro para ENOS. Vê-se que há um desequilíbrio na freqüência de ocorrências de ODP após 1998-1999.

2. OBJETIVO

O objetivo geral deste estudo é investigar a relação entre as interações oceano-atmosféricas e as flutuações das massas de gelo nos Andes tropicais usando dados de sensoriamento remoto e parâmetros meteorológicos. Isto poderá ser obtido através dos seguintes objetivos secundários:

- Calcular as variações nas altitudes da linha de neve (ALN) das geleiras andinas nos trópicos internos e trópicos externos usando sensoriamento remoto durante 1985–2011
- Estimar a relação entre a ocorrência de El Niño - Oscilação Sul (ENOS) e a variação nos ALNs das geleiras
- Calcular as anomalias de temperatura e precipitação durante o período de estudo
- Compreender o comportamento diferencial se houver, entre as geleiras nos trópicos internos e externos com variabilidade de ENOS e anomalias de temperatura e precipitação
- Estimar os efeitos da Oscilação Decenal do Pacífico (ODP) com as variações de ALN durante ENOS

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CAPITULO II

Revisão Bibliográfica

Glacier monitoring and glacier-climate interactions in the tropical Andes: a review

Veettil, B.K.; Wang, S.; Bremer, U.F.; Souza, S.F.; Simões, J.C.

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Glacier monitoring and glacier-climate interactions in the tropical Andes: a review

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Abstract: In this review, we summarized the evolution of glacier monitoring in the tropical Andes during the last few decades, particularly after the development of remote sensing and photogrammetry. Advantages and limitations of glacier mapping, applied so far, in Venezuela, Colombia, Ecuador, Peru and Bolivia are discussed in detail. The applicability of various methods to use glacier records to understand and reconstruct the Andean climate since the Last Glacial Maximum is also explored in this paper. Glacier parameters such as the equilibrium line altitude, snowline and mass balance were given special attention in understanding the complex cryosphere-climate interactions, particularly using remote sensing techniques. Glaciers in the inner and the outer tropics were considered separately based on the precipitation and temperature conditions within a new framework. Results from various studies published recently were analyzed and we tried to understand the differences in the magnitudes of glacier responses towards the climatic perturbations in the inner tropics and the outer tropics. Inner tropical glaciers, particularly those in Venezuela and Colombia near the January Intertropical Convergence Zone (ITCZ), are more vulnerable to increase in temperature. Outer tropical glaciers respond to precipitation variability very rapidly in comparison with the temperature variability, particularly when moving towards the subtropics. We also analyzed the gradients in glacier response to climate change from the Pacific coast towards the Amazon Basin as well as with the elevation. Based on the results from recent studies, it is hypothesized that the glaciers in the inner tropics above the January ITCZ and the southern wet outer tropics will disappear first as a response to global warming whereas glaciers in the northern wet outer tropics and dry outer tropics show resistance to warming trends due to the occurrence of cold phases of El Niño-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) together. Mountain glaciers in Ecuador show less retreat in response to the warming trend, probably due to high altitudes (above 5750 m), in comparison to glaciers in Colombia and Venezuela. However, elevation-dependent warming (EDW) is a major concern in the tropical Andes. In a nutshell, smaller glaciers at lower altitudes in the inner tropics and the southern wet outer tropics near the Amazon Basin are disappearing faster than other glaciers in the tropical Andes.

1. Introduction

Glacier records such as moraine chronology or mass balance time series contain valuable information about the climate of the past (Leclercq and Oerlemans, 2012). Tropical glaciers are one of the most visible indicators of climate change due to their fast response time, sensitivity to climate variations (Hastenrath, 1994), proximity of melting conditions (Kaser and Osmaston, 2002), and clear visibility of their reaction (loss or gain of mass balance) (Lemke et al., 2007; Vuille et al., 2008a). Tropical glaciers differ from other glaciers due to its geometry of the incoming solar radiation as well as the seasonal regime of ablation/accumulation. The accumulation of snow occurs at higher altitudes (and low temperature) while ablation continues at the terminus (lower altitude and higher temperature) (Chevallier et al., 2011).The definition "tropical" is applied only if glaciers are situated within the three boundaries; i.e. within the tropics of Cancer and Capricorn, Intertropical Convergence Zone (ITCZ), and where diurnal temperature range equals annual temperature range.

Glacier-climate interactions in the tropics can be influenced by external forcings such as solar irradiance modulation (Jomelli et al., 2009; Rabatel et al., 2006, 2008), volcanic activities (Angelis et al 2003; Veettil et al., 2016a, 2016b), and the presence of light-absorbing particles such as black carbon on glacier surfaces (Schmitt et al., 2015). An increase in the incoming solar radiation during less cloudy days can influence glacier mass balance sensitivity towards a change in other meteorological variables such as air temperature (Rabatel et al., 2006), even though longwave radiation can provide energy for melting on high-albedo surfaces in mountain environments under cloudy skies (Sicart et al., 2010). Glacial maximum in Peru and Bolivia occurred during the Maunder Solar Minimum (1645-1725) and glacier retreat in the tropics was observed to be slowed down during the Dalton Solar Minimum (1783-1830) (Rabatel et al., 2005a), even though many glaciers in the Alps were retreated or no advances were reported during the Maunder Minimum (Luterbacher et al., 2001; Zasdani, 2007). Volcanic activities influence mass balance changes both directly (Ginot et al., 2010) and indirectly (Rabatel et al., 2013, Veettil et al., 2016a, 2016b), and may be able to mask the influence of climate change on glaciers. Direct influence can be due to glaciochemical activities of ions (Ginot et al., 2010) by altering refreezing temperature levels or reduction in the surface albedo due to particulate materials. Indirect influence includes reduction in the solar irradiance due to volcanic aerosols such as sulphates in the stratosphere. A well-known example for the second type of complex phenomenon is the eruption of Mount Pinatubo in Philippines (even though this volcano is far from the tropical Andes) which suppressed the prevailing warming conditions during 1992-1995 due to a cooling effect of volcanic aerosols in the stratosphere (Angelis et al 2003; Rabatel et al., 2013; Veettil et al., 2016a, 2016b, 2016e). The presence of light-absorbing particles such as black carbon from human population centres also causes increased surface melting by absorbing energy from solar radiation as well as albedo reduction (Schmitt et al., 2015).

Based on the new Randolph Glacier Inventory (RGI) (Arendt et al., 2012; Pfeffer et al., 2014), more than 95% of all tropical glaciers on the earth are situated in the Andes of South America and the remaining are distributed in Africa and Asia. The Andes extend over varying temperature and precipitation zones and are influenced (mainly) by the Atlantic circulation patterns in the north and influenced highly by the Pacific circulation in the south (Kaser, 2001; Favier et al., 2004a; Rodbell et al., 2009; Sagredo and Lowell, 2012; Veettil et al., 2016a). The mean maximum height of the Andes is more than 4000 m above sea level (a.s.l.) along the tropical and subtropical region. This physical barrier disrupts the atmospheric circulation resulting in contrasting climatic conditions on the western and eastern slopes of the Andes. Tropical Andes can be classified into inner and outer tropics based on the seasonality of precipitation, humidity, and accumulation-ablation conditions.

Climatic conditions in this region favour year round ablation towards the terminus (Francou et al., 2004).

More than 80% of the freshwater supply in the arid and semi-arid regions of the tropics and the sub-tropics originates in the mountain regions which are initially stored as ice in glaciers (Messerli, 2001; Vuille et al., 2008a). The balance between accumulation and ablation of these water buffers can break at any time by perturbations in climate thereby causing an imbalance in the river flow regimes in the nearby basins. Tropical Andean glaciers are important contributors of water resources in many high elevation basins, particularly in Bolivia and Peru, during the dry season (Ribstein et al., 1995; Mark and Seltzer, 2003; Villacis, 2008; Soruco et al., 2009; Chevallier et al., 2011; Gascoin et al., 2011). Many of the catchments near the Cordillera Blanca are glacially fed (Juen et al., 2007; Baraer et al., 2015). Runoff from basins nearby fast retreating glaciers would be higher in the beginning, then decreases and finally decline. Recently, Soruco et al. (2015)observed that run off in La Paz city in Bolivia is sustained by increased melt rates, where nearly 15% of the annual water resources (14% in the wet season and 27% in the dry season) are contributed by glaciers. They estimated, with no change in precipitation, annual runoff may diminish up to 12%, if the glaciers disappear completely.

This review article is an insight to glacier monitoring in the tropical Andes and a literature review on glacier-climate interactions in this region. All the relationships between glaciers and climate mentioned in this paper are regarding clean glaciers and not include the complex mass balance fluctuations shown by debris-covered and rock glaciers in the tropical Andes. In the next section, we discuss the suitability of glaciers in this region as a proxy for understanding the climate since the Pleistocene (2,588,000 to 11,700 years ago) and various methods used so far to understand the glacial history of the tropical Andes. Then we cover the significance of mass balance, equilibrium line altitude (ELA) and other glaciological variables in understanding the cryosphere-climate interactions in this region and the differences in the magnitudes of glacier response towards climate perturbations in the inner and outer tropics within a new framework of classification of Andean glaciers (Sagredo and Lowell, 2012; Sagredo et al., 2014). Subsequent sections will discuss the gradients in climate conditions and glacier response to climate change in the tropical Andes, glacier response to ocean-atmosphere interactions such as El Niño – Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO), and glacier monitoring along the tropical Andes (Venezuela, Colombia, Ecuador, Peru and Bolivia). Finally, a few suggestions for future research were included in the conclusion section.

2. Tropical Andean glaciers as records of the climate in the past: methods and their applicability

This section discusses the application of glaciers in the tropical Andes as a proxy to understand the climate of the past. In the discussion of glacial records of the past, we included glaciation during the Pleistocene which includes the Last Glacial Maximum (LGM) (25,000 to 13,000 years ago) and during the Little Ice Age (LIA) (between 16^{th} and 17^{th} century in South America). Glacier fluctuations must be dated well for a better understanding of past climate (Kelly et al., 2015). Various observations on glacier extent from Ecuador, Bolivia, Peru, Chile, Venezuela, and Colombia demonstrated that many rapid but interrupted glacier shrinkages have occurred since the LIA (Jomelli et al., 2009; Vuille et al., 2008a). Fluctuations in glacier length provide an excellent (and sometimes the only) proxy for reconstructing global and hemispheric temperatures (Leclercq and Oerlemans, 2012) as no reliable meteorological records existed before the 19^{th} century. However, the response of glaciers to climate change can vary with many factors such as altitude, surrounding climate, glacier geometry, orientation, geographical location of the glacier within the continent, and volcanic activities.

Equilibrium line altitude (ELA) is one of the dominant glaciological variables used for understanding the glacier-climate interactions of the past (Jomelli et al., 2009) because it is not influenced by glacier morphology or topographic features of the catchment. Knowledge of the past climate variability in the Andes helps us to understand the present trend in climate and to develop climate models for predicting future changes. General circulation models (GCMs), such as those used for the fourth report of the Intergovernmental Panel on Climate Change (IPCC-AR4), suggested that dry season precipitation may increase by the end of the21^{st} century which would effectively enhance the seasonal hydrological cycle in the tropical Andes (Vuille et al., 2008a). Knowledge of tropical Andean climate in the past has its significance in understanding the global climate because climate variations and glacier fluctuations in this region are controlled by the ocean-atmospheric phenomena such as ENSO, PDO and Antarctic Oscillation (AAO) (Seiler et al., 2013a, 2013b).

Reconstruction of climate in the past based on glaciers can be done by using ice core records (Thompson et al., 1984, 1985, 1995, 2006a; Henderson et al., 1999; Bradley et al., 2003; Hoffmann et al., 2003; Knüsel et al., 2003, 2005; Ramirez et al., 2003; Schotterer et al., 2003; Vuille et al., 2003b; Thompson et al., 2006a; Vimeux et al., 2009) or by analysing glacial moraine deposits (Schubert, 1974; Mahaney et al., 2000a; Stansell et al., 2007; Jomelli et al., 2009). Most of the Andean glaciers were, so far, monitored based on ELA derived from moraine mapping (Clapperton, 1991).Other methods include radiocarbon dating of glacial moraines(Mercer and Palacios, 1977; Clapperton, 1981, Hammen et al., 1981; Gouze et al, 1986; Schubert and Clapperton, 1990; Seltzer, 1992; Seltzer et al., 1995), lichenometry (Rodbell, 1992; Rabatel et al., 2005a;Rabatel et al., 2006; Solomina et al., 2007; Jomelli et al., 2008, 2009), numerical models (Malone et al., 2015), Thermo Luminescence (TL) (Schubert and Vaz, 1987; Bezada, 1990), Optically Stimulated Luminescence (OSL) (Mahaney et al., 2000b), and more recently, Terrestrial Cosmogenic Nuclide (TCN) (Farber et al., 2005; Smith et al., 2005; Wesnousky et al., 2012; Angel et al., 2013, 2016; Carcaillet et al., 2013; Guzmán et al., 2013).

Andean ice cores are unique due to their high altitudinal locations, even though their dating is difficult due to complex seasonal cycles (Vimeux et al., 2009). Various ice core records studied in the tropical Andes were taken from Ecuador (Chimborazo and Antizana), Peru (Nevado Huascaran, Quelccaya ice cap, and Nevado Coropuna) and Bolivia (Nevado Sajama, Pomerape and Illimani). The ice core from Quelccaya Ice Cap in 1983 was the first of its kind from the tropical Andes (Thompson et al., 1984). The result obtained from another Quelccaya ice core in 2003 was similar to the 1983 core with additional data for the intervening years (Thompson et al., 2006a). Tropical Andean ice core dating was, so far, done based on annual layer counting (ALC), beta activity, and isotopes like ^{210}Pb, ^{3}H, ^{14}C and ^{137}Cs. ALC is one of the most accurate methods which is widely used in the Andes except for Nevado Coropuna. The basic principle behind ALC method is the variations in insoluble particles and isotopic composition of ice with seasonal changes (Vimeux et al., 2009).Knüsel et al. (2003) applied a non-destructive technique based on the electrical conductivity measurement (ECM – electrical conductivity method), which is dependent on the differences in ions (mainly H^+) through the layer, for ALC of two ice cores from Nevado Illimani, Bolivia. The accuracy of ALC method can be verified by reference horizons such as volcanic dust trapped in ice cores (Vimeux et al., 2009). For example, the eruption of Mt. Pinatubo (1991) was identified in the ice cores from Nevado Illimani (Angelis et al., 2003). One of the most accurate radioactive element dating of the ice core for short periods is ^{210}Pb (half-life: 22.3 years) whereas the temporal resolution of ^{14}C dating is coarse (half-life: +/- 5730 years). Andean ice cores from Peru (Huascaran and Quelccaya) and Bolivia (Sajama and Illimani) showed coherent trends from a 3-year-scale to decadal climate variability over the last 100 years and thereby allow the construction of an Andean Isotope Index (AII) (Hoffmann et al., 2003). However, isotopic records for the last 250 years are not so coherent compared to the last 100 years (Thompson et al., 2006a), and the definition of a long term AII can be questionable.

Ice core records can provide detailed evidence on the influence of ENSO (Thompson et al., 1984; Henderson et al., 1999; Bradley et al., 2003; Hoffmann et al., 2003; Knüsel et al., 2003, 2005; Herreros et al., 2009; Vimeux et al., 2009; Veettil, 2012; Vuille et al., 2003b), AAO (Vimeux et al., 2009), and PDO (Veettil et al., 2014a, 2016a, 2016b) in the tropical Andes. Bradley et al. (2003) used ice cores from Sajama ice cap in Bolivia to record ENSO events and thereby reconstructing Pacific sea surface temperatures (SSTs) in such a way that stable isotopes ($\delta^{18}O$) tend to be enriched during El Niño events and depleted during La Niña events (Vimeux et al., 2009). Other methods to identify ENSO events from ice cores were based on pollen concentration at Sajama during 1958-1996 (Liu et al., 2007) and the presence of ions at Illimani during 1887-1999 (Knüsel et al., 2005).

It is possible to extend ice core study from seasonal to annual and even to millennial scales and hence they provide excellent proxy records for the past climate. However, ice core records are not free from limitations and the accuracy of dating decreases from the surface to the inner cores which makes them less accurate on a millennial scale compared with centennial or decadal scales. The differences in isotopic diffusion rates in the firn and the underlying ice as well as the effects of temperature on the diffusion gradient from the top to the bottom are still not understood completely (Vimeux et al., 2009). Moreover, many studies using ice core records do not consider the loss of fallen snow due to sublimation and wind erosion, and the slope characteristics of the surface (Bradley et al. 2003; Hoffmann et al., 2003; Thompson et al., 2006a). Hence unaccounted differences may exist between the calculated accumulation and net accumulation (Vimeux et al., 2009).

Analysing glacial moraine deposits can provide information of climate in the past based on isotope studies. Moraine deposits during a glacier advance or retreat can be distinguished and are well preserved in the tropical Andes (Rabatel et al., 2013). For larger timescale analysis of glacier changes, such as since the LGM, glacial moraine deposits have proven to give valuable information (Clapperton, 1983; Schubert and Clapperton, 1990; Jomelli et al., 2009) similar to those from radiocarbon dating techniques. The difficulty in using radiocarbon dating in the Andes originates from its dry climate and sparse vegetation to get organic matter trapped in moraine deposits during glacial advances (Seltzer et al., 1995). Jomelli et al. (2011) used ^{10}Be for dating 57 moraines in the Bolivian Cordillera Real for the past 11,000 years and observed irregular retreating patterns of Telata glacier. Recently, Jomelli et al. (2014) updated the glacier chronology in the Tropical Andes using ^{10}Be moraine chronology to study the major glacial advances during the Antarctic cold reversal (14,500 to 12,900 years ago). Martin et al. (2015) have done an in situ study on cosmogenic ^{10}Be production rate in the same region which is potentially helpful to create accurate chronology of glacier fluctuations in the tropical Andes during the late glacial period. Stroup et al. (2014) and Kelly et al. (2015) also used ^{10}Be moraine dating near Quelccaya Ice Cap in Peru, which can be used in other areas of the tropical Andes, where radiocarbon dating is nearly impossible due to the scarcity of trapped organic materials. Another option is to use ^{18}O isotope, which is used by Quesada et al. (2015) for a study on Sajama Ice Cap (Bolivia) to study the impact of palaeolake evaporation on the isotope enrichment of the glacier during the last glaciation (18,500 to 11,700 years ago).

In the tropical Andes, many glaciers are situated above quaternary volcanoes and the eruptive products can interfere with moraine dating in these regions. One of the limitations of moraine deposits based studies is the time lag between the mass balance fluctuations and the movements of the glacier terminus which can add some uncertainties to the temporal resolution. There are a few non-climatic factors, such as the accumulation area topography and glacier hypsometry, which control the accuracy of moraine dating (Barr and Lowell, 2014) and these factors should be considered before using moraine deposits as a proxy for palaeoclimate. The patterns of moraine deposits are dependent on the slope as well; higher slope increases the velocity of surface materials (Francou et al., 2001). Geomorphic records can be incomplete due to the destruction of older moraine deposits by a new glacier advance (Jomelli et al., 2009).

Lichenometry, using the growth curves of lichens such as *Rhizocarpon sp*. that are commonly growing on glacial moraine deposits, can be used for understanding glacier fluctuations in the tropical Andes. Rabatel et al. (2005a, 2006, and 2008) used the radial growth rate of *Rhizocarpon Geographicum* for dating glacier fluctuations since the Little Ice Age (LIA) in the Cordillera Real in Bolivia. Later, Jomelli et al. (2008) also applied the same methodology to estimate glacial advances during the LIA in Cordillera Blanca, Peru. The basic principle is that the diameter of the lichen growth on a surface is proportional (in some way) to the time since the surface is open to lichen growth under specific environmental and lithological conditions (Rabatel et al., 2006) and then applying this proportionality to another surface, where its age is to be calculated. Lichen growth rates can be different on the dry western slopes of the tropical Andes from the wet eastern slopes (Solomina et al., 2007). Moraine dating using lichenometry was applied since the beginning of the Holocene (Rodbell, 1992) to a few centuries (Rabatel et al., 2005a; Solomina et al., 2007). Instead of radiocarbon deposits, lichenometry can be used widely as they grow under extreme conditions such as cold and high altitudes, where other types of vegetation used to be absent in the past (Rodbell, 1992). Early models such as the one used by Innes (1985) assumed that lichen growth follow normal distribution and the confidence intervals were based on Gaussian distribution. New models, such as those proposed by Cooley et al. (2006) and Naveau et al. (2007), used extreme value distribution and the uncertainty associated with lichen diameter and the corresponding age distribution is reduced (Rabatel et al., 2005a).

In addition to understanding glacier fluctuations, lichenometric analysis add some light to climate variations in the past, such as variations in precipitation, historical ENSO records (Rabatel et al., 2005a) and atmospheric conditions (whether dry or wet and hot or cold), even though lichen growth rates can also be altered due to environmental changes (Innes, 1985). Similar to the case of using moraine deposits, lichen growth curves can be removed or altered by a new glacial advance at the terminus and hence can add some 'incompleteness' in the dating of glaciers. However, if each moraine is considered as a separate study site or a control point as proposed by Solomina et al. (2007) to find the largest lichen in the Cordillera Blanca, this problem can be solved as more samples are taken and thereby eliminating the dependency on the accuracy of just one sample.

Tropical Andean glaciers, from the north to the south, have shown similar trends in response to climatic variations in the past or the palaeoclimatological evidences force us to believe so (Mercer and Palacios, 1977; Clapperton, 1981, Hammen et al., 1981; Gouze et al, 1986; Seltzer, 1992; Seltzer et al., 1995). To reconstruct a long-term climate trend in the tropical Andes from glaciers, a combination of the above-mentioned methods can be helpful. A summary of literature on glacier chronology in the tropical Andes and various methods used by researchers to estimate glaciers changes in the past are summarized in **Table 1**.

Table 1: Summary of studies mentioned in this study and the methodologies used on past glacier changes in the tropical Andes. The list is by country from north to south

3. Equilibrium line altitude, snowline, and mass balance of glaciers in understanding glacier-climate interactions in the tropical Andes

Equilibrium line altitude (ELA) is one of the main glaciological variables used widely in the tropical Andes to understand the influence of climate change on mass balance (Jomelli et al., 2009). The annual ELA delimits the accumulation zone (annual mass balance >0) from the ablation zone (annual mass balance<0) and is approximately equivalent to the annual snowline maximum (Condom et al., 2007; Rabatel et al., 2012) in the tropics, and particularly in the outer tropics. An increase in the ELA denotes a negative mass balance and vice versa. Generally, ELA increases with increase (decrease) in temperature (snowfall).

The traditional definition of ELA refers to the altitude where $b_n = 0$; b_n is the net ablation at the end of ablation season (summer) (Benn et al., 2005). In the case of tropical glaciers, where year-round ablation occurs, this definition is not suitable (Benn et al., 2005) and the ELA can be above the 0°C isotherm. Greene et al. (2002) expressed ELA for the tropics as a function of the atmospheric freezing isotherm and the mean annual precipitation by multiple regression analysis as given below:

$$ELA = 537 + 1.01FH - 0.51P \quad (\textbf{Eq 1})$$

Where ELA is the equilibrium line altitude in meters; FH is the annually averaged freezing height in meters; P is the mean annual precipitation in millimetres. A lapse rate of 6.5°Ckm^{-1} is commonly used in the tropical Andes for inferring the temperature variations from freezing altitude (Jomelli et al., 2009). This definition (**Eq 1**) is not general for the entire tropical Andes. Based on Fox (1993), who calculated ELA of glaciers in Peru and Bolivia from snowline at freezing altitudes and annual precipitation amount, Condom et al. (2007)

established a generalized empirical equation to calculate the ELA between 5°S and 20°S as given below:

$$ELA = 3427\text{-}1148(\log_{10}(P)) + T/0.007+z \qquad \textbf{(Eq 2)}$$

Where the ELA is in m a.s.l., P is the annual precipitation (mm/Yr), T is the mean annual temperature at ground level (°C) and z is the station elevation (m a.s.l.). In general, annual snowline from 10°N to 12°S is situated near 4700 m a.s.l. (ELA is close to the atmospheric 0°C isotherm) and the glacier terminus ranges from 4100 to 5000 m a.s.l. From 12°S to 28°S the annual snowline rises from 4700 m a.s.l. to 6000 m a.s.l. due to dry conditions (around 25°S, the ELA is situated nearly 1000 m above the 0°C isotherm) (Condom et al., 2007).

Polissar et al. (2006) calculated a 1,500-year glacier reconstruction model and uncertainties in temperature estimation based on ELA in Venezuela during AD 500-2000. Rabatel et al. (2006), based on mean ELA variations, reconstructed the glacier recession on Cerro Charquini in the central Andes, Bolivia between the LIA and 1997. It is possible to calculate ELA of tropical Andean glaciers (or alpine glaciers in general) on millennial timescale using statistical approaches and readers are requested to refer Seltzer (1994) for a detailed description. Though a complete record is not available, interpreting the variations in glacier length or ELA is commonly used during the last few decades (Benn et al., 2005; Jomelli et al., 2009). However, field data is necessary to calculate ELA and this is not possible in the case of many high altitude glaciers in the tropical Andes.

Rabatel et al. (2012) used an indirect method, based on annual snowline from satellite data, to calculate the nearest ELA (during the dry austral winter) of the year in the outer tropics. For mid-latitude glaciers, the highest snowline at the end of the hydrological year can be taken as a representative of the annual ELA (Lliboutry, 1965) and this representativeness was later validated in the French Alps (Rabatel et al., 2005b) and the tropical Andes (Rabatel et al., 2012; Veettil et al., 2014a, 2016a, 2016b, 2016e, 2016f). However, seasonal snowfall may cause underestimation of the ELA from snowline and the applicability of snowline as an indicator of the ELA is not homogenous from the inner to the outer tropics. Broecker (1997) argued that the measured tropical snowline may vary with the sea level rise, and this type of uncertainty should be corrected accordingly, though such errors are or not so significant.

Other methods to estimate the vertical profile of mass balance and ELA (glacier elevation index – GEI, in some cases) in mountain environments include the accumulation area ratio (AAR) which is the fraction of a glacier surface that has net accumulation (Barry, 2006), balance ratios (BR), terminus-to-headwall altitude ratios (THAR), and Toe-to-summit altitude method (TSAM) (Benn et al., 2005). AAR and BR methods can be applied only when accurate topographic maps are available; whereas THAR and TSAM can be applied in remote regions without topographic data (Benn and Lehmkuhl, 2000). The vertical profile of mass balance in the tropical Andes is important because there is a large vertical gradient in mass balance in the ablation zones, which is influenced by climatic settings. Ablation gradients tend to be steeper in the wet tropics compared with dry tropics (Benn et al., 2005). In order to improve the reliability of results, Benn and Lehmkuhl (2000) recommended ELA reconstruction from multiple methods in order to get a clear picture of the mass balance characteristic of tropical glaciers.

When dealing with the mass balance of tropical Andean glaciers, it is important to note that:

1. Precipitation (and hence accumulation) occurs during the wet summer months whereas ablation occurs year-round and hence these glaciers are known as 'summer accumulation type'
2. Sublimation has an important contribution, particularly during dry winter and evapo-sublimation dominates the ablation process during this period.
3. The magnitude of mass balance depends on the relative position of the glacier with the 0°C isotherm.

Because tropical glaciers are considered as reliable indicators of climate change, it is important to consider the factors that influence mass balance and how faster the glacier response is. For modeling glacier variations with climate change and the interpretation of climate in the past using glacier fluctuations as a proxy, it is important to consider the lag between perturbations in climate and glacier changes. Many factors can influence the glacier response time to climate change such as topography and geometry (Raper and Braithwaite, 2009; Zekollari and Huybrechts, 2015), local climate conditions, and the amount of solar radiation. Glacier mass loss in humid environments tends to be faster because of their increased sensitivity towards temperature (Oerlemans and Fortuin, 1992). Mass balance fluctuations in the tropical Andes depend on cloud cover that alters long-wave radiation budget and seasonal changes in the precipitation (whether snow, sleet or rain). However, mass balance sensitivity towards long-term climate conditions depends not only on climate conditions (temperature, precipitation, humidity, surface radiation etc.) but also on the surface albedo characteristics (Klok and Oerlemans, 2004), elevation, exposure (Soruco et al., 2009), slope, and surface area or size (Veettil et al., 2016b). Various climate signals recovered or estimated in the tropical Andes, so far, from glacier records include temperature (Klein et al., 1999; Vuille et al., 2003b; Braithwaite et al., 2006; Stansell et al., 2007; Brauning, 2009), precipitation (Thompson et al., 1985; Klein et al., 1999; Vuille et al., 2003b) and humidity (Ammann et al., 2001; Brauning, 2009).

Many researchers studied the complex relationship between glacier mass balance variability in the tropical Andes and large-scale circulation (Arnaud et al., 2001; Bradley et al., 2003; Favier et al., 2004a, 2004b; Francou et al., 2004; Knüsel et al., 2005; Liu et al., 2007; Vuille et al., 2008a, 2008b; Veettil, 2012; Veettil et al., 2014a, 2016a, 2016b and many more). Most of these studies correlated fluctuations in glaciological variables such as ELA, snowline and ice core records with ENSO, PDO and Sea Surface Temperature Anomalies (SSTAs). A new study (Maussion et al., 2015) showed that the relationship between glacier mass balance and ENSO is stronger than that observed in previous studies (Arnaud et al., 2001; Francou et al., 2004; Vuille et al., 2008b; Veettil et al., 2014a). Even though PDO signatures on climate variability in the tropical Andes are well known (Seiler et al., 2013a, 2013b), its relationship with glacier mass balance is relatively a new topic (Veettil et al., 2014a, 2016a, 2016b). Studies using ice cores on PDO signals in glacier records are not available in the tropical Andes as done by Gao et al. (2015) near the Tibetan Plateau.

4. Andean glaciers in the inner and outer tropics and the new framework to classify tropical Andean glaciers

The Andes span different climatic zones with varying temperature, humidity, and precipitation conditions (Veettil et al., 2016b) with higher influence of the Atlantic circulation in the north, the Pacific circulation influence in the south (Sagredo and Lowell, 2012), and a combined influence of both the Pacific and the Atlantic in the Central Andes. However, it is reported recently that the north Pacific also influences the Venezuelan Andes in the north (Polissar et al., 2013). The surface energy balance (SEB) measurements (Favier et al., 2004a) and the surface energy and mass balance (SEMB) model (Rupper and Roe, 2008; Sagredo et al., 2014) have provided valuable information on glacier behaviour in the tropical and the subtropical climate (**Figure 1**). In general, the tropical Andes can be divided in to two subzones based mainly on precipitation conditions and geographical location (Rabatel et al., 2013). Firstly, the inner tropical Andes (Venezuela, Colombia and Ecuador)with continuous precipitation throughout the year with year-round ablation and accumulation, and secondly, the outer tropical Andes (Peru, Bolivia and northern Chile), where dry season occurs from May to September and a wet season from October to March (Rabatel et al., 2013) with notable accumulation only during the wet season (Jomelli et al., 2009).Both the inner and the outer tropics maintain homogenous temperature conditions throughout the year and glaciers in both regions are subjected to higher daily temperature variations than annual temperature variation. However, a small seasonality in air temperature (1^{o}C - 2^{o}C difference between the austral summer and the austral winter) is found in the outer tropics (Rabatel et al., 2013) and this makes this region somewhat similar to the subtropics. In other words, in the outer tropics, the climate conditions are 'tropical' only during the wet austral summer (Kaser and Osmaston, 2002) and subtropical conditions prevail during the dry austral winter whereas permanently humid conditions prevail in the inner tropics.

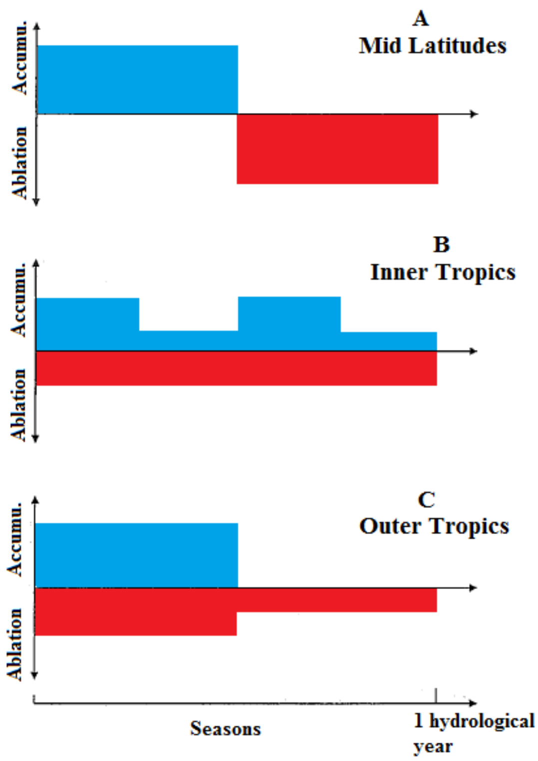

Figure 1: Ablation characteristics of glaciers in mid-latitudes, inner tropics and outer tropics (Modified from Kaser and Osmaston, 2002)

There exists a seasonality in ablation and accumulation of glaciers in the mid and high latitude glaciers with seasonality in moisture (Kaser, 1996a, 1996b). In the outer tropics, mass loss (melting and sublimation) occurs throughout the year whereas accumulation is limited to the precipitation season (austral summer). In both inner and outer tropics, the differences in annual precipitation and annual cloudiness are not so significant but the seasonal distributions of these two variables are entirely different (Favier et al., 2004a). The inner tropics exhibit little variations in air temperature and 0°C isotherm altitude. Precipitation occurs throughout the year but due to the presence of the ITCZ, there exists some seasonality with two slight maxima in precipitation per year (Kaser, 1995a; Sagredo et al., 2014). The ITCZ is a belt of low pressure and intense low-level convergence of trade winds over the equatorial oceans (Garreaud et al., 2009). During the austral winter (JJA), maximum precipitation occurs towards the north of the equator almost along the oceanic ITCZ and at the same time, the central part of the continent is devoid of any precipitation. During the austral summer (DJF), a rapid southward shift of convection occurs and heavy precipitation occurs from the southern half of the Amazon Basin towards the north of Argentina. During the austral fall, the ITCZ returns gradually towards the north of the equator.

Mass balance of glaciers in the inner tropics, where the 0°C isotherm changes in the ablation zone, is more sensitive to temperature changes compared with the outer tropics, and hence interannual variations in temperature influence the ablation characteristics (Francou et al., 2004). This explains why the glaciers in Colombia and Venezuela are disappearing faster than ever, under global warming (Morris et al., 2006). Precipitation on glaciers is not always solid in the inner tropics and the precipitation phase (whether snow or rain) influences surface albedo that plays an important role in mass balance changes. Higher rate of snowfall increases the surface albedo and thereby reduces melting. The opposite is true during rain and hence both the temperature and the precipitation phase control the annual mass balance in the inner tropics. The situation is slightly different in the outer tropics, where rainfall rarely occurs in the high elevation ablation zone and a small seasonality in temperature exists. The precipitation variability and sublimation plays an important role in determining the annual mass balance of glaciers in the outer tropics.

Using the SEB measurements, despite being situated in two relatively different climatic conditions, Favier et al. (2004a) observed a common response from the glaciers of both the inner and the outer tropics towards a few climatic perturbations such as ENSO. Similar responses to ENSO in the inner and outer tropics, where glaciers are subjected to different climatic conditions, originate from their temperature and precipitation sensitivities, respectively. During El Niño, air temperature increases and being more sensitive to temperature, glaciers in the inner tropics undergo enhanced ablation. On the other hand, glaciers in the outer tropics are having a 4-to-6 months of cold and dry season with reduced melting (Favier et al., 2004a), and a precipitation deficit induced by the El Niño causes the glaciers in the outer tropics to undergo negative mass balance. However, for low altitude glaciers in the outer tropics, a precipitation deficit is not necessary for higher ablation because a rise in the air temperature enhances rainfall instead of snow and thereby reduces the surface albedo.

4.1. Tropical Andean glaciers within the new framework

Based on statistical analysis of three climatic variables - temperature (annual mean, annual range and 0°C isotherm elevation), precipitation (total annual, annual range and seasonality), and humidity (annual mean and annual range) - and a selected number of glaciers with simple geometry, Sagredo and Lowell (2012) classified the Andes into seven climatic groups and three groups among them belong to the tropical Andes. These three groups are: inner tropics (Venezuela, Colombia and Ecuador), wet outer tropics (Northern and central Peru and eastern cordilleras of Bolivia) and dry outer tropics (southern Peru, western Cordilleras of northern Chile and Bolivia). The wet outer tropics can be divided into two groups based on temperature seasonality and average humidity (**Figure 2**).The first group (northern wet outer tropics), north of 13°S, has uniform seasonality in temperature (near 0°C with average humidity 71%) and covers the Peruvian Western Cordillera (including the Cordillera Blanca). The second group (southern wet outer tropics), south of 13°S, covers the Eastern Cordillera of Peru and Bolivia and has an oscillating temperatures even up to or more than 4°C throughout the year, even though the humidity is lower (approximately 60%) than the first subgroup. Precipitation cycle in the southern wetter tropics is more complex and other than ITCZ shift, the precipitation is controlled by the South American summer monsoon as well (Sagredo et al., 2014).

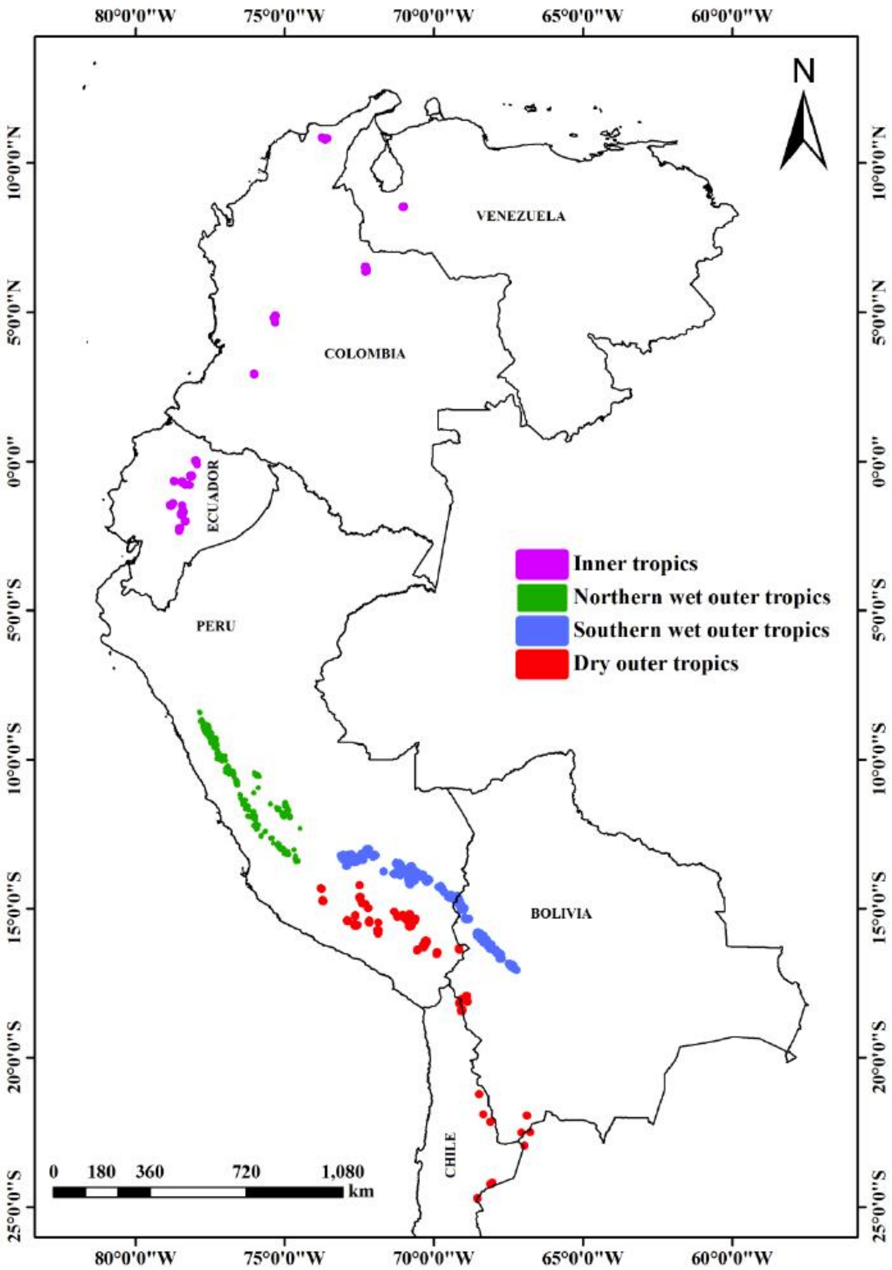

Figure 2: Glaciers in the tropical Andes classified based on temperature, precipitation and humidity conditions

Using the above mentioned classification and SEMB model (Rupper and Roe, 2008), Sagredo et al. (2014) estimated the sensitivity of ELA to precipitation and temperature along the Andes. They used three glacier surface variables (albedo, surface roughness and snow emissivity) along with meteorological variables (precipitation, wind speed, relative humidity, atmospheric pressure, net incoming shortwave radiation, temperature lapse rate, air temperature and atmospheric emissivity) using CRU CL 2.0 data from Climate Research Unit

(CRU), University of East Anglia and NCEP/NCAR Reanalysis data. The estimation of ELA in this model was done using terminus-to-head altitude ratio (THAR) and used a single THAR value for the entire Andes. ELA of Glaciers in the inner tropics and the wet outer tropics are described to be highly sensitive to global warming (Sagredo and Lowell, 2012) by shifting the summer 0°C isotherm towards higher altitudes whereas a precipitation deficit would cause higher glacier retreat in the dry outer tropics (Favier et al., 2004a). Note that the glaciers in the inner tropics are identical in the annual precipitation cycle with the glaciers in the Cordillera Darwin in Chile at the southern end of the continent, and it is the annual temperature range that differentiates the glaciers in the Cordillera Darwin (7.4°C) from those glaciers in the inner tropics (1.1°C) (Sagredo and Lowell, 2012). Generally, glaciers in the wet outer tropics are summer-accumulation type. The dry outer tropics, which cover the glaciers in the western cordilleras of southern Peru, Bolivia and northern Chile, experience extremely cold and dry conditions (average humidity nearly 50%) and the glaciers in this region are located above the mean annual 0°C isotherm. Between the southern wet outer tropics and the dry outer tropics, a transition from temperature influence to precipitation influence on the glacier ablation is visible. At the same time, a sublimation dominated glacier ablation (common in the subtropics) prevails over the melt dominated ablation (which is found in the inner and wet outer tropics) during this transition. However, both sublimation and melting were found to have significant contribution in glacier ablation in the northern wet outer tropics (Maussion et al., 2015).

Even though we used the above-mentioned framework of classification (Sagredo and Lowell, 2012) to discuss the glacier-climate interactions along the tropical Andes, their methodology and assumptions are not free from flaws. The number of samples (glaciers in this case) considered for cluster analysis is relatively small (234 glaciers), even though their samples represented all glacier groups and the results were compared and refined with principal component analysis (PCA) and analysis of similarities (ANOSIM). The number of samples taken affects the results of the classification (Lai and Huang, 1989). This method can be refined using relatively complete glacier inventories such as the RGI. Another issue is with the smaller size of glaciers (approximately < 2 km) with simple geometry used; mass balance changes in response to climate change is not independent on size (Rabatel et al., 2013) and geometry (Diolaiuti et al., 2011). THAR value is dependent on the shape of the glacier and varies in each group and because single THAR value is selected for the entire Andes for evaluating ELA sensitivities (Sagredo et al., 2014), it contributes some errors.

Asymmetry in the number of glaciers in each group may also contribute errors. For example, inner tropics are having smaller number of glaciers and are classified as a single group whereas the warming trend in Venezuela and Colombia (see section 7.1.2) is higher compared with that in Ecuador (Veettil et al., 2014a). Considering the glacier variables such as albedo along the Andes as a constant (0.62 in this model) for estimating ELA sensitivity to temperature and precipitation (Sagredo et al., 2014) using the SEMB model also add some discrepancies because these variables vary from place to place as well as seasonally. These discrepancies can be reduced by using different surface albedo values at different climate zones (inner tropics, outer tropics, subtropics etc.). Another criticism may arise be from the coarse resolution or sparse coverage of the meteorological datasets used. CRU CL 2.0 data used in this classification has a higher latitude/longitude resolution (10´) but is not free from interpolation errors (New et al., 2002) due to sparse coverage in high mountain areas. Resolution of the NCEP/NCAR Reanalysis data (2.5°x2.5°) is not too high to account for the specificities of each group, particularly where wet-to-dry transition occurs. High resolution data can be used in the future to overcome these limitations.

Despite the above mentioned limitations, this framework is still better than generalising the tropical Andes just as inner and outer tropics to describe the complex cryosphere-climate interactions along the tropical Andes. It provides a first approximation of the climate at each site, as the authors hypothesised.

5. Gradients in climate conditions and glacier response to climate change in the tropical Andes

Even though the glacier response to climate change is qualitatively similar, the magnitude of response varies with climatic regimes of glaciers (Kaser, 2001; Favier et al., 2004a, 2004b; Fujita, 2008a, 2008b; Rupper and Roe, 2009; Sagredo and Lowell, 2012; Veettil et al, 2014a, 2016a, 2016b). As discussed in the previous section, seasonal variations in temperature are limited in the tropics and seasonal changes in the environment are primarily due to the variations in precipitation, humidity and moisture advection (Jomelli et al., 2009). In general, the SEB is controlled by net short-wave radiation (S) in the tropics and S is controlled by surface albedo and cloud cover (Favier et al., 2004a, 2004b). This means that cloud cover controls short-wave radiation incident on glacier surfaces whereas albedo controls the portion of the incident radiation energy available for ablation. A variation in the ELA is an immediate response of the mass balance budget towards a changing climate

(Kaser, 1995a). Many researchers hypothesized that, glaciers in general, are more sensitive to changes in temperature than to changes in precipitation (Giesen and Oerlemans, 2010; Leclercq and Oerlemans, 2012). It is seen that air temperature in the tropical Andes has increased by 0.13°C/decade between 1960 and 2010 and then there is a recent hiatus, particularly in the coastal region (Vuille et al., 2015). A recent study (Veettil et al., 2016e) in the Cordillera Blanca observed that, despite the hiatus, the decrease in the glacier area was higher during 2005-2015 and that warming still continues at higher elevations. Glaciers in wet climate are more sensitive to temperature variations and those glaciers in dry areas respond relatively slow towards a warming trend (Rapper and Braithwaite, 2009; Rupper and Roe, 2009). A study (Veettil et al., 2016a) on the Nevado Coropuna in the Cordillera Ampato (Peru) showed that the above mentioned statement is not valid always, at least during the 1997/98 El Niño, where precipitation variability was little to nonexistent and yet there was glacier retreat. There was a small increase in precipitation in the inner tropics during the second half of the 20th century and a minute decrease in the same was observed in the outer tropics (Vuille et al., 2008a). Moreover, glaciers above the 0°C level are highly sensitive to precipitation variations and insensitive to temperature variations (Kaser and Osmaston, 2002; Sagredo and Lowell, 2012).

The climate of the South American continent and the atmospheric circulation systems are influenced by the frontal systems throughout the year. Freezing temperatures and enhanced precipitation are expected during these events at its trajectory (Parmenter, 1976), particularly over the Llanos (5° – 10° N) (Poveda et al., 2006). The frequency of occurrences of cold fronts in the austral winter (JJA) is maximum and in the austral summer (DJF) is minimum. Barret et al. (2009) observed that the precipitation patterns in the region during cold fronts are modified by the presence of the Andes because this topographical barrier is an obstacle to the advancement of cold fronts (particularly the low-tropospheric flow).Vuille and Amman (1996) observed that the enhanced snowfall events in the dry outer tropics are mainly due to the moisture coming from the Pacific during cold front events.

5.1. West-to-east gradient in climate conditions and glacier response in the tropical Andes

The tropical Andes experience both east-west and north-south climate gradients and the west-to-east climate gradient is due to the separation of low-level circulation (Garreaud et al., 2009). The west of the Andes maintains dry and cold conditions (and hence moisture transport from the west to the Andes is low) whereas the eastern parts are very humid due to the moisture transport from the tropical Atlantic and hence high precipitation occurs over the Amazon basin. The Andes are a formidable obstacle for the tropospheric circulation and foster tropical-extratropical interactions along the east (Poveda et al., 2006; Garreaud et al., 2009) and are responsible for the asymmetries in precipitation and temperature in the tropical and subtropical South America. ENSO phenomenon originates in the tropical Pacific and has strong direct influence on the regional climate and glaciers on the western sides of the tropical Andes (Veettil, 2012) with or without in a combined mode with PDO (Veettil et al., 2014a, 2016a, 2016b). At the same time, Atlantic SST variations influence the climate on the eastern sides of the tropical Andes. In tropical Andes, monthly mean precipitation increases from the Pacific coast towards the Amazon Basin (Veettil et al., 2014a, 2016a, 2016b) and the precipitation pattern (not the amount) is almost similar from the Pacific towards the east. The Eastern and the western slopes of the Ecuadorian and Colombian Andes have precipitation fed by the moist air from the Amazon Basin and the Gulf of Panama respectively (Bendix et al., 2006). However, precipitation along the eastern slopes is higher due to the moisture content carried by the mid-level easterly flow from the Amazon Basin (Garreaud, 2009). A local maximum in the precipitation along the west of the Andes is caused by the uplift of low-level winds by the western slopes whereas the forced subsidence over the eastern sides of the Andes produces dry conditions in the subtropics (Garreaud et al., 2009). Maximum west-to-east contrast in precipitation is observed between 18ºS and 23ºS (Garreaud, 2009) with very cold and dry Atacama Desert (average elevation: 4,000 m a.s.l.) in the west and the Chaco wetlands (average elevation: 310 m a.s.l.) in the east, even though the elevation is not so high in this region. Other than the asymmetries in precipitation, there exists a west-to-east asymmetry in temperature as well: relatively cold (and dry/arid) conditions prevail along the Pacific coast whereas warm (and humid/rainy) conditions prevail in the inner parts of the continent along the eastern slopes. Urrutia and Vuille (2009) showed that below 1000 m a.s.l., the eastern side is warmer than the western side, the western side is warmer between 1000 m a.s.l. and 3000 m a.s.l., and the eastern side is warmer again above 3000 m a.s.l.

Many recent studies observed that the glaciers situated in the southern wet outer tropics (eastern cordilleras of Peru and Bolivia) are retreating faster than ever (Ramirez et al., 2001; Ribeiro and Simões, 2010; Veettil et al., 2016b, 2016e, 2016f) compared with those glaciers in the western cordilleras (Maussion et al., 2015; Veettil et al., 2016b). One of the reasons behind this anomaly is due to the exceptional increase in relative humidity (Vuille et al., 2008a), which is transported from the Amazon Basin to the Andes. An increase in the atmospheric humidity reduces sublimation and enhance melting (mass loss due to sublimation is less compared with melting) due to the reduction in the vapor pressure difference between ice surface and air (Wagnon et al., 1999). Another factor is the precipitation phase –eastern cordilleras of Peru and Bolivia is supplied with higher rate of rain, which reduces surface albedo and increase melting. Moreover, those glaciers in the western cordilleras were observed to have no or reduced retreat during La Niña periods (Arnaud et al., 2001; Veettil et al., 2014; Maussion et al., 2015) whereas the influence of La Niña in the eastern cordilleras are less (Veettil et al., 2016b).

5.2. North-to-south gradient in climate conditions and glacier response in the tropical Andes

Both upper-level and low-level flow patterns vary from the north to the south in the Andes. Upper-level circulation is characterized by moderate easterly winds at low latitudes (15ºN-15ºS) and by westerly winds towards the subtropical region. Low-level flow is more complex than upper-level circulation and is responsible for the moisture transport, thereby controlling the precipitation field (Garreaud, 2009). A north-south gradient in the tropical Andean climate is more visible from the precipitation patterns. Precipitation in the tropical Andes is due to the deep convective storms developing over the mountain range and the chance of precipitation in the form of snowfall increases from the north of the tropical Andes (Colombia and Venezuela) towards the south. In the northern tropical Andes (inner tropics), snowfall occurs only on the highest peaks of the mountains and the snowfall elevation decreases towards the subtropics. The inner tropics (Colombia and Ecuador) are not having a specific single precipitation season whereas the outer tropics (Peru, Bolivia and northern Chile) are characterized by specific dry and wet seasons (**Figure 3**). Within the outer tropics, rainy season occurs during November-April in Peru and only during December-March in Bolivia and northern Chile (Garreaud et al., 2003). In the inner tropics, precipitation in Ecuador is bimodal – precipitation occurs during March-May and September-November (Vuille et al., 2000a) whereas the Colombian Andes is very humid and the topography modulates the precipitation with main wet season during June-August (Poveda et al., 2001). ENSO and other global phenomena influence the tropical Andean climate differently from the north to the south and this means that glacier response to such phenomena also varies from the north to the south. In the tropical Andes, the northern most part (of 0.5ºN) is the only region where the temperature variability is correlated significantly with the North Atlantic SST anomalies (Vuille et al., 2000a) whereas the rest follow the SST anomalies in the Nino-3 and Nino-3.4 regions with a lag of 1-3 months (Kumar and Hoerling, 2003). Along the Andes, from the north to the south, the Altiplano has its own climate conditions with low temperature and air pressure with high radiative input and extremely dry conditions (except the austral summer) because of zonal wind anomalies. There exists a north-to-south gradient in the summer climate of the Altiplano itself: northern half is very humid compared with the southern half (Garreaud, 2009). Summertime circulation over South America is characterized by the establishment of the upper level Bolivian High, an upper-level anticyclone centred at 17°S/70°W, which is induced by the deep convection over Amazon Basin (Garreaud et al., 2009).

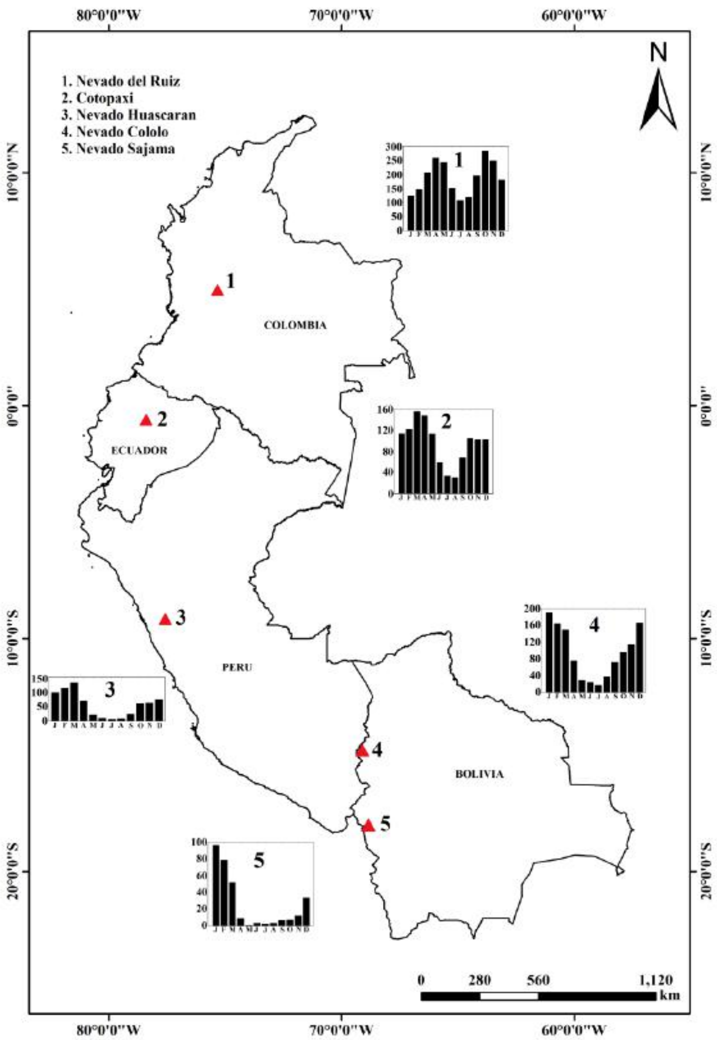

Figure 3: Variations on monthly mean precipitation along the tropical Andes (obtained from NCEP-NCAR Reanalysis data for the period 1950-2008)

Glacier retreat in the tropical Andes from the inner tropics to the outer tropics is not uniform. The extinction of glaciers in the inner tropics (particularly in Colombia and Venezuela) closely follows global warming trends (Rabatel et al., 2013; Veettil et al., 2016b). It is observed recently that the glacier retreat in the southern tropical Andes (south of cordillera Blanca) is not as high as the retreat observed in the northern regions (north of Cordillera Blanca) (Veettil et al., 2016a, 2016b). North-to-south gradients in glacier behaviour in the tropical Andes are already covered in section 4.

5.3. Elevation gradient in climate conditions and glacier response in the tropical Andes

The climate of high altitudes is unique in many aspects such as precipitation patterns, temperature, air pressure, humidity, and solar radiation. It is known that many glaciers at very high altitudes, such as those in the Karakoram Himalaya, are showing a zero or even a positive mass balance (Hewitt, 2005; Veettil, 2012; Veettil et al., 2014b, 2016c) and low altitude glaciers in the same geographical region are not showing any advance and some of them are even retreating (Fujita and Nuimura, 2011; Veettil et al., 2014b; 2016c, 2016d). This shows that elevation is an important factor that controls mass loss, even though the lower troposphere warming is described to be elevation independent (Oerlemans, 2005). On the other hand, a few studies based on general circulation models, Bradley et al. (2006) for example, hypothesized that the rate of warming increases with altitude. This type of elevation dependent warming (EDW) can affect high altitude tropical Andean glaciers when minimum temperature values show stronger tendency towards EDW (Pepin et al., 2015). There are a few studies showing a general decreasing trend in warming with elevation (Vuille and Bradley, 2000; Rangwala et al., 2009; Lu et al., 2010) and these observations can be dependent on the time periods examined, number of station data used, elevation range, and temporal resolution of the datasets used (Pepin et al., 2015). Many factors such as the complex snow-albedo feedback mechanism (Pepin et al., 2015), cloud cover (Liu et al., 2009), water vapour and latent heat release (Rangwala et al., 2009; Rangwala, 2013), soil moisture, atmospheric aerosols such as black carbon, and a combination of all these factors (Pepin et al., 2015) are identified as the possible causes of EDW. Regional variability in EDW can be influenced by large-scale circulation variability such as ENSO and PDO (Diaz et al., 2014) and there are no global relationships between elevation and warming rates (Pepin and Lundquist, 2008). Lack of reliable datasets at very high altitudes in the tropical Andes (approximately above 4000 m a.s.l.) is a real problem in validating such models. For a better understanding of EDW, efforts could be focussed on measuring those factors that influence this phenomenon (snow, cloud, moisture, etc.) and the sparseness of a few *in situ* instruments (land surface temperature measurement, for example) can be replaced by satellite remote sensing (Pepin et al., 2015).

Despite the EDW, many high altitude glaciers in Ecuador (Veettil et al., 2014a), Peru (Veettil et al., 2016a) and Bolivia (Veettil et al., 2016b) are undergoing relatively slower retreat compared with those glaciers in Colombia and Venezuela (Schubert, 1998; Ceballos et al., 2006; Braun and Bezeda, 2011). Some models (Urrutia and Vuille, 2009) showed a decrease in mean annual surface temperature with increase in the elevation on both western and eastern slopes of the tropical Andes and the changes were more abrupt on the western sides. Another study (Vuille and Bradley, 2000) also calculated similar trends in temperature with a pattern of reduced warming in the tropical Andes with increase in elevation, particularly on the western slopes and temperature trends on the eastern slopes increases up to an elevation of 2500 m a.s.l. with no change up to 1000 m a.s.l.

Precipitation changes along the eastern and western slopes are more complex and the mass balance depends on the accumulation of snow at the upper parts and the ablation at the lower parts. Since solid precipitation (snow) prevails at very high altitudes (above 5,500 m a.s.l., for example), though moderated by flow conditions (slope, for example), it is obvious that the elevation has a positive feedback on the mass balance and orographic effects are decisive in glacier response to climate change, which is more intense in the outer tropics. However, both incoming solar radiation and seasonal temperatures vary with elevation. *In situ* observations in the Ecuadorian Andes show that precipitation increases with increase in elevation (Garreaud, 2009; Veettil et al., 2014a). This means that depending on the region (inner tropics, wet outer tropics or dry outer tropics); the decisive elevation for sustaining a positive mass balance can be different. It is observed that the inner tropical glaciers in Colombia and Venezuela have retreated much faster during the last few decades (Schubert, 1998; Ceballos et al., 2006; Braun and Bezeda, 2011), particularly after the PDO regime shift in 1976 whereas those glaciers in Ecuador have not undergone such a fast retreat (Veettil, 2012; Veettil et al., 2014a). Concisely, elevation dependency on the influence of climate change on tropical Andean glaciers needs further investigation using some reliable meteorological datasets at high altitudes (say 4000 m a.s.l., for example).

6. Modern glacier changes in the tropical Andes and interannual/interdecadal variability in the Pacific Ocean

In this section, modern glacier fluctuations (since the LIA) in the tropical Andes related to the influence of climate variability and ENSO/PDO-related climate variability are discussed. In the tropical Andes, LIA was identified during the 16th-19th centuries (Kaser, 1999; Rabatel et al., 2005a) and most of the glaciers in this region reached their LIA maximum in the 17th century (Rabatel et al., 2005a). Unlike those glaciers in the European Alps, no LIA advances were reported in the tropical Andes in the early 19th century (Rabatel et al., 2005a). On interannual and interdecadal timescales, the climate of the tropical Andes and particularly in the Altiplano, is found to be linked to tropical Pacific SST anomalies (Vuille et al., 2000a; Garreaud and Aceituno, 2001; Francou et al., 2003) and the glacier mass balance is directly linked with the atmospheric conditions in the middle of the austral summer (DJF). Various ocean-atmospheric phenomena, such as ENSO, PDO, and AAO, affect the Andean climate (Seiler et al., 2013a, 2013b) and the relationship between mass balance and precipitation, mass balance and temperature, and mass balance and humidity can be interrelated. In the last five decades, there was a warming of 0.1^{o}C/decade in the tropical Andes and a significant warming of 0.2^{o}C/decade was observed during the last three decades (Vuille and Bradley, 2000; Vuille et al., 2003) (roughly concurrent with the so called "Pacific shift" or the PDO phase change from its cold regime to warm regime in 1976). Vuille et al. (2003) also mentioned that the eastern slopes underwent weaker warming compared with the western slopes, which are directed towards the Pacific. Moreover, the northeastern Andes are dominated by ENSO during December-February whereas in the eastern Cordillera, this is during June-August (Vuille et al., 2000a) and below average precipitation occurs in both regions during the El Niño episodes. Numerous studies observed variations in temperature and precipitation during the occurrences of ENSO (Aceituno and Garreaud, 1995; Poveda and Mesa., 1997; Vuille et al., 2000a, 2000b; Garreaud and Aceituno, 2001; Garreaud, 2009, Garreaud et al., 2009) and PDO (Andreoli and Kayano, 2005; Garreaud et al., 2009; Seiler et al., 2013a, 2013b) in the tropical Andes.

6.1. ENSO-related climate variability and glacier changes in the tropical Andes

ENSO is one of the most discussed phenomena on the influence of interannual variability on glaciers in the tropical Andes. The annual variations in precipitation (and temperature) in the tropical Andes are caused primarily due to the modification of the Walker circulation (Tadeschi and Collins, 2015) by ENSO, a coupled ocean-atmosphere phenomenon with warm (El Niño) and cold (La Niña) episodes along the equatorial Pacific. The influence of ENSO on mass balance is primarily due to the influence on precipitation. The interannual glacier changes on a larger spatial scale are controlled by the SST anomalies in the eastern equatorial Pacific (Nino 2+1 region) (Francou et al., 2003). Ribstein et al. (1995) observed higher runoff from the Cordillera Real during El Niño periods. Describing the relationship between precipitation variability and mass balance fluctuations during ENSO must be done cautiously. During the strong El Niño in 1997/98, as reported by Herreros et al. (2009), there was not much variation in precipitation near the Nevado Coropuna in Peru. During the same period, there was an increase in snowline (Veettil et al., 2016a) and this shows that the glaciers in the outer tropics fluctuate during the occurrence of ENSO, even if the rate of precipitation remains unchanged. However, the frequency of occurrence and the time of apparition of snowfall must be taken into account because they influence the melting process. Glaciers in the tropical Andes (mainly in the outer tropics) experience a precipitation deficit and hence albedo reduction during El Niño events, along with higher air temperature and reduced cloud cover (that prevents shortwave radiations under normal conditions). In addition to near surface warming, humidity levels were also increased recently in the tropical Andes (Vuille et al., 2003; Veettil et al., 2016e) that influence the sublimation, precipitation, and cloudiness.

During El Niño, a strong westerly flows over the tropical Andes that inhibits the moisture transport from the Amazon basin towards west and this causes a reduced precipitation in the tropical Andes (Bolivia and southern Peru, in particular) (Vuille and Keimig, 2004) and a warmer than normal air temperature along the tropical and the subtropical latitudes (temperature variations of up to 1°C were observed during ENSO) which results in a negative mass balance of glaciers in the tropical Andes (Francou et al., 2004). A decrease in precipitation over the northern Andes (inner tropics) during El Niño episodes due to relaxed land-sea thermal contrast and enhanced subsidence by deep convection over the eastern Pacific ITCZ (Poveda et al., 2001) can cause extreme dry conditions. In the outer tropics (including the central Andes), there exists a dry condition (not 'extreme' as in the inner tropics) due to the decrease in the advection of moist air from the continent originated from strong mid-level westerly flow. In general, below normal precipitation and "warmer than normal" conditions were observed in the tropical and subtropical Andes during El Niño episodes. Moreover, opposite conditions were found during the La Niña episodes (Vuille et al., 2000a, 2000b) and the impact of La Niña is stronger than El Niño's impact (Poveda and

Mesa, 1997). Maximum correlation between ENSO and precipitation in the Altiplano was observed during December-February and between ENSO and temperature was observed during December-March in the tropical Andes (Garreaud et al., 2009).

The variations in the air temperature, as discussed previously, differs on the eastern and the western slopes of the Andes and hence slightly different mass balance variations are expected in the tropical Andes on the two sides. Vuille et al. (2000a, 2000b) and Kumar and Hoerling (2003) showed that there is a lag of 1-3 months between the tropical Pacific SST anomalies and the temperature anomalies in the tropical Andes and a difference of 0.7°-1.3°C in the temperature anomalies in this region. This lag is expected to be reflected on the mass balance variability due to ENSO and must be considered while drawing a conclusion on the ENSO influence on glaciers in this region (Veettil, 2012; Veettil et al., 2014a, 2016a, 2016b). The observed ENSO influence on temperature trends strongly depends on the time period analyzed because of the recent hiatus on global warming (Vuille et al., 2015) (which is related to the observed cold PDO fingerprints) and is likely to be reflected on tropical Andean temperature (Schauwecker et al., 2014). A detailed review of the seasonal correlation between the Multivariate ENSO Index and precipitation/air temperature is described in Garreaud (2009).

One of the first papers on the influence of ENSO on Andean glaciers based on remote sensing datasets, Arnaud et al. (2001), described how the annual snowline of Nevado Sajama in Bolivia varied between 1963 and 1998 using aerial photographs and Landsat images. The increase in the frequency of occurrence and stronger intensity (above the +2.0°C SST anomaly is considered as very strong intensity El Niño) of El Niño events during the period between 1991 and 2001 was hypothesised to be linked directly with the recent disappearance of glaciers such as Chacaltaya in the Cordillera Real in Bolivia (Francou et al., 2003). The strong El Niño events during 1982/83 and 1997/98 (strongest in the last six decades) showed a high correlation between ENSO indices and mass balance variations in the tropical Andes. However, the prolonged El Niño condition during 1991-1995 was observed to be not associated with a negative mass balance as expected (Veettil et al., 2016a, 2016b) and this anomaly is assumed to be linked with stratospheric aerosol clouds followed by the Mt Pinatubo eruption in 1991 (Rabatel et al., 2013, Veettil et al., 2016a, 2016b). ENSO influence on glaciers in the tropical Andes should be considered with caution because the ENSO-related climate variability in South America is dependent on PDO phases (Andreoli and Kayano, 2005), which is discussed in the next section.

6.2. PDO-related climate variability and glacier changes in the tropical Andes

PDO and interdecadal pacific oscillation (IPO) are considered as ENSO-like phenomena because their spatial climate fingerprints during the warm and cold phases are similar to the El Niño and the La Niña events, respectively. However, it is suggested to consider ENSO and PDO for research in the Southern Hemisphere because IPO is more associated with ENSO (Pezza et al., 2007). Unlike ENSO, PDO can persist from a few years to several decades and is the leading principal component of the monthly SST anomalies in the north Pacific (Mantua et al., 1997). PDO primarily influences the north Pacific whereas its effects reach near the equator. The anomalous SST pattern related to PDO is nearly symmetric about the equator but is less equatorially confined in the eastern Pacific (Silva et al., 2011). Moreover, the Sea Level Pressure (SLP) shows a strong annular structure related to PDO during the austral summer, which is not seen for ENSO index (Pezza et al., 2007; Silva et al., 2011). However, PDO may not be considered as totally independent of ENSO, and both of them yield wetter subtropics and drier tropics over the south and north Americas. Shakun and Shaman (2009) considered PDO as a reddened response to ENSO forcing from the tropics. The PDO-related anomalies in precipitation and temperature are similar to ENSO-related anomalies but smaller in magnitude (Garreaud et al., 2009). Andreoli and Kayano (2005) and Kayano and Andreoli (2007) verified that the intensities of ENSO-related teleconnection are higher (lower) when ENSO and PDO are in (opposite) phase. ENSO and PDO are the sources of climate variability in Bolivia along with AAO (Seiler et al., 2013a; Seiler et al., 2013b). It is interesting to note that the rainfall variability in Bolivia roughly follows PDO patterns with an increase (of 12% in December-February and 18% in June-August) during 1965-1984 and a decrease (of 4% in December-February and 10% in June-August) during 1985-2004 (Seiler et al., 2013b). Some researchers (Ebbesmeyer et al., 1991; Garreaud et al., 2009) consider the PDO phase change in 1976 as a “Pacific climate shift”. However, PDO is not the only cause of this shift because the frequency of El Niño events also increased since the 1975. For example, there were no El Niño events that can be considered as very strong (above +2.0^{o}C SST anomaly) observed during 1950-1975 whereas three such events occurred during 1976-2016. There was only one moderate El Niño (1.0 to 1.4^{o}C SST anomaly) during 1950-1976 and five such Niño events occurred during 1976-2010.

Veettil et al. (2014a, 2016a, 2016b) demonstrated a combined influence of PDO and ENSO on the snowline altitudes of mountain glaciers in the inner and the outer tropics and at the same time Wang et al. (2014) observed that this combined influence occurs on the global land dry-wet changes as well. Veettil et al. (2014a) showed that ENSO influence the snowline variations in the tropical Andes mainly when El Niño occurs during the positive PDO or when La Niña events occur in phase with negative PDO. PDO entered a prolonged positive regime since the late 1970s and the frequency of occurrences of El Niño events are increased recently; this would explain the recent snowline rise of glaciers in the tropical Andes. In a recent by Veettil et al. (2016e), a chronology of glacier fluctuations in the Cordillera Blanca since the early 14th century and the occurrences of phase changes of PDO and ENSO during this period were presented. It is observed in this study that earlier glacial advancement during 1330+/-29 estimated by Solomina et al. (2007) occurred during the cold phase of PDO reconstructed by McDonald and Case (2005) and the maximum glacial extent during 1630-1680 also coincides with PDO cold phase. Many short-lived glacier advances occurred in the Cordillera Blanca around mid-1920s (Ames, 1998; Kaser, 1999) towards the end of a prolonged PDO cold regime. Most of these advances before the 20th century occurred during stronger or long-lived La Niña phases, based on the reconstructed ENSO indices by Khider et al. (2011) and Yan et al. (2011), coupled with PDO cold phases. However, these are merely observations and the limitations of reconstructed ENSO and PDO indices prevent us from drawing a conclusion without further investigation.

7. Glacier monitoring along the tropical Andes within a new framework

This section discusses various studies on glacier monitoring in the tropical Andes and the subsections are organized in such a way that glaciers in each of the four climatic groups described in section 5 (inner tropics, northern wet outer tropics, southern wet outer tropics and dry outer tropics) are discussed separately. Special attention is given to recent studies using remote sensing and photogrammetric techniques. Details of the representatives of glaciers considered studied using remote sensing and aerial photographs in each group discussed in this paper are summarized in **Table 2** and their relative geographical locations are given in **Figure 4**. Regional meteorological and hydrological datasets, including river discharge data, in each country are maintained by the corresponding national organizations as given below:

Venezuela: Instituto Nacional de Meteorología e Hidrología (INAMEH)

Colombia: Instituto de Hidrología, Meteorología y Estudios Ambientales (IDEAM)

Ecuador: El Instituto Nacional de Meteorología e Hidrología (INAMHI)

Peru: Servicio Nacional de Meteorologia e Hidrologia (SENAMHI)

Autoridad Nacional del Agua del Perú (ANA)

Bolivia: Servicio Nacional de Meteorologia e Hidrologia del Bolívia (SENAMHI)

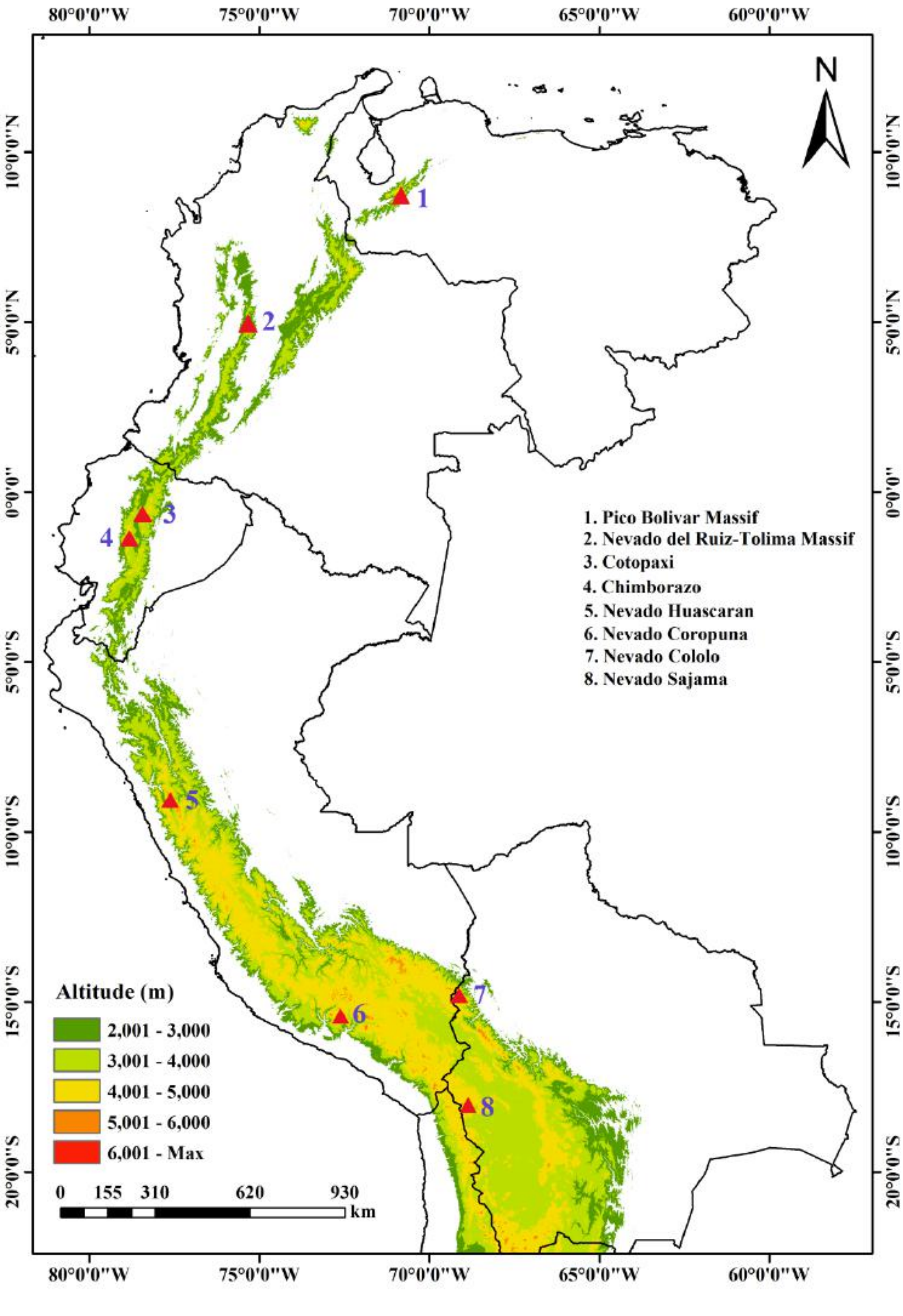

Figure 4: Locations of various glaciers in the tropical Andes mentioned in this paper

Table 2: Details of satellite data/maps used in glacier mapping/mass balance determination studies in the topical Andes described in this review

7.1. Inner tropics

Inner tropical Andes are situated in Venezuela, Colombia, and Ecuador and majority of the inner tropical glaciers are situated in Ecuador. Glaciers in Venezuela and Colombia are already near extinction and those in Ecuador are retreating relatively slower compared with other glaciers in the inner tropics.

7.1.1. Venezuela

Glaciers in Venezuela and Colombia were affected by the quaternary glaciations (Morris et al., 2006) and the history of glacier fluctuations in Venezuela is similar to elsewhere in the tropical Andes (Morris et al., 2006; Braun and Bezeda, 2013). Limited number of studies exists on the modern glacier fluctuations in Venezuela (Jahn, 1925; Schubert, 1984, 1992, 1998; Schubert and Clapperton, 1990; Morris et al., 2006; Carillo and Yépez, 2008; Braun and Bezeda, 2013). There was about 200 km^2 of glaciated area in the entire Sierra Nevada de Merida during the Last Glacial Maximum, out of which the Pico Humboldt-Pico Bonpland Massif alone constituted about 140km^2 (Schubert, 1984). However, the first mapping of modern glaciers during 1910-1911 in Venezuela was done by Jahn (1925) and he mapped a glaciated area of 10 km^2 in three massifs namely the Pico Bolivar, the Pico La Concha and the Pico Humboldt-Pico Bonpland (7 km^2 in the Pico Humboldt-Pico Bonpland Massif alone). Jahn (1925) introduced glacier nomenclature to the Venezuelan glaciers and some of them were later renamed by Schubert (1992, 1998) (**Table 3**). Schubert (1992) used aerial photographs for the first time for mapping glaciers in Venezuela in 1952 and calculated that the Sierra Nevada de Merida was having a glaciated area of 2.91 km^2 (Pico Humboldt-Pico Bonpland Massif alone had 2.03 km^2). Satellite images for mapping glaciers in Venezuela is used by Schubert (1998)for the first time and Morris et al. (2006) reported shrinkage of Venezuelan glaciers from 2.03 km^2 in 1952 to 0.20 km^2 in 2003 by combining the results from Schubert (1998) and using ASTER images. Again, in 2011, Braun and Bezeda (2013) claimed that only 0.1 km^2 of glaciated area left in Venezuela (Humboldt Glacier associated with Pico Humboldt, previously described as Laguna Verde Glacier by Jahn in 1925). Based on SPOT-5 images, in 2008, this glacier had an area of 0.33 km^2 (Carillo and Yépez, 2008). The Humboldt Glacier is not having an accumulation area at present and its existence is due to the particular topographic factors, and hence the interpretation of the equilibrium line of this glacier cannot be considered as an indicator of climate change (Braun and Bezeda, 2013). Braun and Bezeda (2013) also mentioned that the glacier retreat in Venezuela after 1972 shows directional properties in such a way that those glaciers facing south and southeast of the Sierra Nevada de Merida are directly exposed to sun in the morning and are retreating faster than those directed towards the north and northwest (as the afternoons in this region are cloudy). This feature combined with the presence of shading effects by the rock ridges on its sides explains the existence of the Humboldt Glacier.

Table 3: Glacier nomenclature in three massifs in Venezuela given by Jahn (1925) and later modified by Schubert (1992, 1998) (in brackets)

It is interesting to see that there was no notable changes in the glacier area in Venezuela between 1952 (Schubert, 1992) and 1976 (Schubert, 1998) and afterwards underwent a rapid retreat. This stability during 1952-1976 might be partially due to the prolonged cold regime of PDO during 1956-1975 along with stronger La Niña events compared with El Niño events as described in Veettil (2014a, 2016a, 2016b).Being smaller in size, Venezuelan glaciers showed relatively fast response to climate fluctuations compared with larger ones in the other parts of the tropical Andes. Unfortunately, we do not have aerial photographs or field data records on an annual basis to test the existence of a correlation between the glacier fluctuations and the PDO phase changes during this period. Moreover, most of the recent optical satellite images in this region are extensively could-covered. The altitude of the 0ºC isotherm of the Humboldt Glacier changed from 4895 m a.s.l. in 1976 to above 5000 m a.s.l. in 2010 (Braun and Bezeda, 2013). In fact, this increase in freezing height was observed in the entire American cordilleras (Diaz et al., 2003a, 2003b) and occurred during the PDO regime shift from cold phase to warm phase in the late 1970s.

7.1.2. Colombia

One of the main objectives of glacier monitoring in Colombia is to prevent hazards due to the occurrences of complex volcano-glacier activities rather than climate change impacts (Naranjo et al., 1986; Ceballos et al., 2006; Huggel et al., 2007; Granados et al., 2015), particularly after the 1985 Nevado del Ruiz/Armero catastrophe (Naranjo et al., 1986). However, the results of the above mentioned studies showed that the glacier retreat in the recent decades is linked directly with the changes in regional and global climate. Many studies exist on recent glacier changes and future predictions in Colombia (Florez, 1992; Collazos, 2002; Pabón, 2003; Ceballos et al., 2006; Morris et al., 2006; Huggel et al., 2007; Herrera and Ruiz, 2009; Poveda and Pineda, 2009), mainly using remote sensing. The Quaternary glaciation history in Colombia is well-documented (Hammen et al., 1981; Herd, 1982; Clapperton, 1983; Hammen, 1985; Helmans and Kuhry, 1986; Helmans, 1988; Schubert and Clapperton, 1990 and many more). In a study using aerial photographs and Landsat images, Ceballos et al (2006) calculated that the glaciers in Colombia lost about 50% of their area during the last five decades and the highest retreat occurred during 1986-2000. An area loss of 32% during 1950-2003 was reported by Morris et al. (2006) on three glaciated regions (the Sierra Nevada de Santa Marta, the Sierra Nevada del Cocuy, and the Ruiz-Tolima Massif) using Aerial photographs, ASTER and Landsat images. Small differences arise in the area calculation depend on the type of images used. Based on a study on the Nevada del Cocuy, Herrera and Ruiz (2009) showed similar results and claimed that the glacier retreat during 1986-2007 was linear and the glacier will disappear within 20 years, if this rate continues. Poveda and Pineda (2009) predicted similar changes by 2020 in their assessment using Landsat images and their calculated period of glacier disappearance is nearly 100 years earlier than as predicted by the 2007 IPCC Fourth Assessment Report. The 2014 IPCC Fifth Assessment Report is in agreement with the results from Ceballos et al. (2006) and Poveda and Pineda (2009) and these results are included in the report. Currently, in Colombia, glaciers mainly exist in the Sierra Nevada de Santa Marta (5775 m a.s.l.), the Sierra Nevada del Cocuy (5490 m a.s.l.) and other four ice-covered volcanoes namely Nevado del Ruiz (5400 m a.s.l.), Nevado de Santa Isabel (5110 m a.s.l.), Nevado del Tolima (5280 m a.s.l.), and Nevado de Huila (5655 m a.s.l.). A satellite based observation of Nevado de Santa Isabel, using Landsat images acquired during the precipitation minimum (JJA), is shown in **Figure 5**.

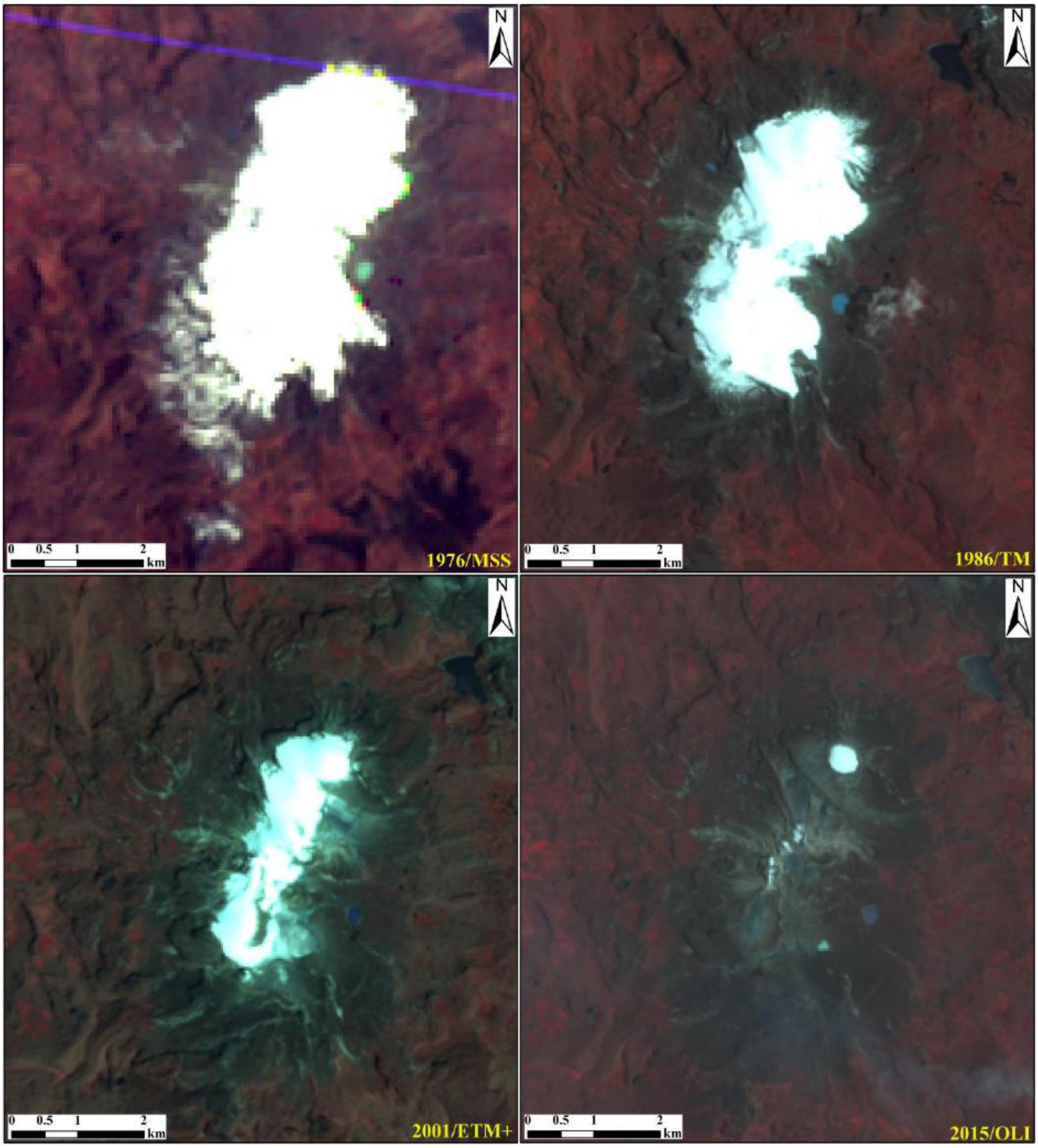

Figure 5: Retreat of Nevado de Santa Isabel during 1976-2015 using Landsat time series

The existence of two precipitation seasons (March-May and September-November) (Refer **Figure 3**) makes the estimation of snowline and equilibrium line of mountain glaciers in Colombia a bit difficult on an annual basis (Veettil et al., 2014a) because the snowline can vary during any (or both) of the two precipitation maxima. It is recommended to use decadal changes in area as a glaciological variable using remote sensing datasets for understanding mass balance changes in this region. Based on meteorological data from high altitudes, Pabón (2003) and Ceballos et al. (2006) noticed that there was an increase of nearly 1-1.5°C in the air temperature during 1970-2000 in Colombia and this might have provoked the rapid shrinkage of glaciers in this region. As expected, smaller glaciers such as the Volcán Nevado del Tolima are retreating faster due to their mid-latitudinal, tropical nature (Francou et al., 2003; Paul et al., 2004) whereas the larger ones such as the Nevado del Ruiz and the Nevado del Huila were experienced a relatively smaller retreat (Ceballos et al., 2006). However, the recent retreat of the highest glacier peak in Colombia, Sierra Nevada de Santa Marta (5775 m a.s.l.), was not as small as expected on a high altitude glacier; there was a decrease (in the glacier area) of about 50% in this region. This anomalous behaviour might be partially due to the climatic regime (<50 km from the Caribbean coast) different from that of other glaciated regions (Ceballos et al., 2006).We have calculated the anomalies in temperature and precipitation near the Nevado del Ruiz using NCEP/NCAR Reanalysis 1 data (Kalnay et al., 1996) and is observed that the air temperature in Colombia shows an exceptionally increasing trend at regional scales, which was not observed in Ecuador, Peru, and Bolivia (Veettil et al, 2014a, 2016a, 2016b, 2016e).

Without considering precipitation anomalies, theorizing the glacier-climate interactions based only on temperature anomalies would be inadequate because the mass balance sensitivity towards precipitation variability and seasonality is not a constant along the tropical Andes (Favier et al., 2004a; Rabatel et al., 2013). The precipitation variability in the tropical Andes is related to the occurrence of ENSO and the Colombian Andes is not an exception for this. The strong El Niño occurred in 1997/98 induced a negative mass balance due to higher temperature and reduced precipitation followed by a slow retreat during the strong La Niña conditions in 1999 (Ceballos et al., 2006). However, in the case of Colombian glaciers, an increase in the air temperature and solar radiation were the principal mechanisms behind the rapid glacier retreat (Collazos, 2002) rather than the changes in precipitation. This assumption must be validated by considering the fact that an increase in temperature can trigger a change in the precipitation phase from solid (snowfall) to liquid (rain) which in turn favours glacier retreat due to albedo reduction. Even though difficult to achieve, it is possible to differentiate rain and snowfall using recent advancements in remote sensing. For example, Wulf et al. (2016) separated the contribution of rain (using TRMM data) and snowfall and glacier melt components (using MODIS data) to river discharge in the western Himalaya. Variations in the rate of precipitation can also be associated with the ENSO and this makes understanding of the glacier-climate interaction in this region more complex.

7.1.3. Ecuador

About 4% of the World´s tropical glaciers are situated in Ecuador and are located in the two mountain chains (cordilleras) – Cordillera Occidental in the west and Cordillera Oriental in the east. In the Cordillera Oriental, some of the lower elevation mountains are glaciated due the direct exposure to the moisture supply from the Amazon basin (Vuille et al., 2008a) whereas only high altitude mountains in the Cordillera Occidental are glaciated. Ecuador had extensive glaciations since the LIA with its maximum extent during the 1730s and later there was a discontinuous recession with smaller advances in 1800, 1850, and 1870 (Hastenrath, 1981, Jomelli et al., 2009, Vuille et al., 2008a). Based on lichenometry, Jomelli et al. (2009) estimated that the ELA of glaciers in Ecuador increased by 200-300 m from the 18th century to the end of the 20th century. The ice-covered Cotopaxi volcano (5,897 m a.s.l.) (Jordan et al., 2005; Cáceres, 2010; Veettil et al., 2014a) and the Antisana (5,704 m a.s.l.) (Francou et al., 2004) in the eastern Cordillera of Ecuador were studied extensively over the past few decades. Glaciated mountains such as Chimborazo (6,310 m a.s.l., Cordillera Occidental) underwent four generations of glacial deposits coupled with volcanic activities (Smith et al., 2005a). Since most of the glaciers in Ecuador are situated on active volcanoes, there exists a danger related to the rapid ice melting due to laharitic phenomenon (pyroclastic material flow from erupted volcanoes). Some of the ice-covered volcanoes have erupted recently, such as the Tungurahua (in 2006, 2008, 2010, 2011, 2012) and Sangay (in 2012) (Veettil, 2012) and frequent eruptions may influence accuracy of the results obtained from ice core records due to ash and acid deposits, even though stable isotopes remain unaffected (Ginot et al., 2010). Cotopaxi is reported to have undergone eruptions during 1741-1744, 1768 and 1977 (Jordan et al., 2005) and recently, it is undergoing eruption since August 2015.Cáceres et al. (2004) tried to quantify the possible rapid ice discharge after a possible eruption of Mt Cotopaxi.

Some of the previously glaciated mountains such as the Cotacachi has already lost their glacial coverage completely and the rest are retreating faster than ever since the early 1980s (Collet, 2010). During 1998-1999, the total glaciarized area in Ecuador was about 97.21 km^2 (21.92 km^2 in the Cordillera Occidental and 75.29 km^2 in the Cordillera Oriental) (Jordan and Hastenrath, 1999). Veettil (2012), using Landsat images, estimated that approximately 80 km^2 (which is about 18% reduction from 1999) of glaciers still exist in Ecuador in 2010-2011. This calculation is still consistent with the 40% reduction in the Ecuadorian glaciers over 40 years as calculated by Cáceres (2010) on the glaciated areas of the Cotopaxi, Antisana, and Chimborazo. The Antizana 15 was reported to be advanced for the period of 1999-2001 during a persistent La Niña phase (Francou et al., 2004). However, a recent study on Antisana 15α (Bastantes-Serrano et al., 2016), using additional geodetic approach, reported a slightly negative mass balance change during the period compared with the results from the previous study (Francou et al., 2004) which used the glaciological approach. Jordan et al. (2005), based on photogrammetry and remote sensing, reported an area loss of about 30% during 1976-1999 with a stagnant phase during 1956-1976. The small advance of the Antizana 15 during 1999-2001 can be attributed to the short negative PDO during 1998-2002 La Niña (Veettil et al., 2014a). The Chimborazo also lost 35% of its glaciated area between 1986 and 2010 (La Frenierre, 2012) and the Antisana lost about 33% during 1979-2009 (Rabatel et al., 2013). In a recent study, Veettil et al. (2014a) observed that when the positive regime of PDO and El Niño occurs in phase, the rate of retreat is higher due to the high temperature sensitivity of these glaciers. Cadier et al. (2007) observed that El Niño takes about four months to influence glacier melting in Ecuador.

7.2. Wet outer tropics

Glaciers in the wet outer tropics are situated in the western cordilleras (including the Cordillera Blanca) in the north of 13°S of Peru and the eastern cordilleras of Peru (such as the Cordilleras Urubamba, Carabaya and Vilcanota) and Bolivia (Cordilleras Apolobamba, Real and Tres Cruces). Various studies estimated different warming rates for different cordilleras in the outer tropics – 0.30°C/decade in the Cordillera Blanca (Mark and Seltzer, 2005); 0.2°C/decade in the Cordillera Huayhuash (McFadden et al., 2011) and the Cordillera Vilcanota (Salzmann et al., 2013) and a general increase of 0.1°C/decade in the tropical Andes during 1939-2006 (Vuille et al., 2008a, 2008b). Based on precipitation conditions, wet outer tropics are again classified as the northern wet outer tropics that are entirely situated in the western cordilleras of northern Peru (north of 13°S) and as the southern wet outer tropics that are situated in the eastern cordilleras of Bolivia and Peru (south of 13°S) (Sagredo et al., 2014). In the northern wet outer tropics, the precipitation pattern is controlled by the southward shift of the ITCZ whereas more complex circulation patterns occur in the southern wet outer tropics due to the ITCZ movement as well as the presence of the Bolivian High and the low-level jet (Chaco Low) (Garreaud, 2009; Garreaud et al., 2009). Generally, glaciers in the wet outer tropics are less temperature sensitive compared with glaciers in the inner tropics and more temperature sensitive compared with those in the dry outer tropics (Favier et al., 2004a; Sagredo and Lowell, 2012).

7.2.1. Northern wet outer tropics

The northern wet outer tropics entirely belong to Peru and the seasonality in temperature is uniform in this region (near 0°C). Glaciers in the Cordilleras Blanca, Huallanca, Huayhuash, Raura and La Viuda belong to this group. Most of the glaciers in Peru belong to Cordillera Blanca, which is a part of the Cordillera Occidental, and are the best documented in the tropical Andes. Mark and Seltzer (2005) observed an average increase of 0.26°C/decade in the air temperature in the Cordillera Blanca during 1961-1999 and that the glacier retreats during this period was asymmetric (ELA changes varied between 25 m and 125 m for different glaciers). Wet and dry periods occur during November-March and May-August respectively, with transitory periods in April and September-October (**Figure 3**). A new study by Schauwecker et al. (2014), using station data from SENAMHI (Peru) and NCEP/NCAR data, proposed a "spatially heterogeneous and temporally discontinuous" changes in temperature and precipitation patterns in the Cordillera Blanca; in this study, the authors observed a higher warming rate (0.31°C/decade) during 1969-1998 and a lower warming trend (0.13°C/decade) during 1983-2012, probably due to the recent hiatus in global warming (Vuille et al., 2015).

Maximum glacier extents in the Cordillera Blanca occurred during 1330-1360 (Solomina et al., 2007) and 1630-1680 (Jomelli et al., 2009). Equilibrium line of glaciers in this region increased up to 100 m between the early 17th century and the early 20th century (Vuille et al., 2008a; Jomelli et al., 2009). There are three small advances reported in this region in the mid-1920s (McLaughlin, 1924; Kaser, 1999), in the late 1970s (Kaser and Georges, 1997), and recently at the end of the 20thcentury (Georges, 2004), which is consistent with PDO phase changes. The ice coverage of the Cordillera Blanca at the end of the 20th century is calculated as slightly below 600 km^2 which was about 800-900 km^2 in the early 20thcentury (Georges, 2004). However, this advance was not a general trend and was not observed in the case of the Huascaran-Chopicalqui and the Artesonraju (Vuille et al., 2008a). There was nearly a 7% loss of glaciated area in the entire Cordillera Blanca between 1987 and 1997 and 15% loss between 1970 and 1996 (Silverio and Jaquet, 2005). Racoviteanu et al. (2008) calculated a loss of 22.4% of the initial glaciated area in the Cordillera Blanca between 1970 and 2003 using aerial photography and SPOT images and noticed that the retreat of debris-covered glaciers was smaller when compared with clean glaciers. However, in contrast to the observation done by Kaser and Georges (1997) during 1930-1950, Racoviteanu et al. (2008) observed higher rate of retreat on the eastern slopes of the Cordillera Blanca. Baraer et al. (2015) calculated the area of four glaciers namely the Llanganuco, the Quilcahuanca, the Yanamarey, and the Pumapampa as 63.76 km^2, 87.66 km^2, 26.93 km^2 and 52.65 km^2, respectively, and the median elevation calculated was between 4946 and 5231 m a.s.l., using the 2005 GLIMS database. The glaciated area of Artesonraju decreased by 20% between 1960 and 2000 based on the GLIMS database (Raup et al., 2006). Ames (1998) reported a retreat of 1079 m on the terminal position of the Broggi Glacier between 1932 and 1987, 690 m on the Glacier Pucaranra and 675 m on the Uruashraju between 1936 and 1994 and 552 m on the Yanamarey Glacier between 1932 and 1994. The retreat of the Yanamarey Glacier was later calculated to be 20 m/year during 1977-2003 (Vuille et al., 2008a). Recently, Burns and Nolin (2014) observed that the decrease in the area of glaciers in the Cordillera Blanca during 2004-2010 was three times more than that during 1970-2003. This trend is in agreement with the snowline changes calculated by Veettil et al. (2016e) in the southern end of Cordillera Blanca, where Alarcón et al. (2015) observed an area loss close to 1.4 km^2per decade between 1957 and 2010.

A recent study by Maussion et al. (2015) confirmed ENSO influence on the surface energy balance and mass balance in the Cordillera Blanca based on a study conducted on the Shallap Glacier and also observed that the precipitation/temperature variability is the driving mechanism rather than sublimation during the occurrence of ENSO. There was an exceptional increase in relative humidity, slightly higher than what expected from the observed temperature increase, in the southern Cordillera Blanca (Veettil et al., 2016e) and this may enhance melting instead of sublimation in this region due to the reduction in the vapour pressure difference between snow and air (Wagnon et al., 1999).

Glaciers in the cordilleras Huallanca, Raura, La Viuda and Huayhuash are subjected to the same climate conditions as in the Cordillera Blanca. McFadden et al. (2011) calculated the variations in annual snowlines of selected glaciers in the Cordillera Huayhuash and the Cordillera Raura using Landsat images. The average snowline of the Cordillera Huayhuash changed from 5062+/-36 m a.s.l. to 5086+/-33 m a.s.l. during 1986-2005 and that of the Cordillera Raura increased from 4947+/-7 m a.s.l. to 5044+/-8 m a.s.l. during 1986-2002 (McFadden et al., 2011). These temporal changes in the mean snowline display some potential differences between these two cordilleras; snowline variation in the Cordillera Huayhuash was statistically insignificant compared with that in the Cordillera Raura and this can be partially attributed to the differences in elevation and initial size. In general, the increase in snowline of glaciers on the western slopes were faster, compared with those on the eastern sides and similar trends were observed in the Cordillera Blanca by Kaser and Georges (1997). The increasing trend in SLA in the Cordillera Raura can be explained by the linear the increase in temperature (0.1^{o}C to 0.11^{o}C per decade) in the tropical Andes as calculated by Vuille and Bradley (2000), even though this model does not fit to explain the case of Cordillera Huayhuash (McFadden et al., 2011).

7.2.2. Southern wet outer tropics

Other than glaciers in the Cordillera Blanca in the northern wet outer tropics, some of the glaciers in the southern wet outer tropics, such as the Quelccaya Ice Cap (Thompson et al., 1979, Thompson, 1980, Albert, 2002; Salzmann et al., 2013; Hanshaw and Bookhagen, 2014), Nevado Illimani (Liu et al., 2013; Ribeiro et al. 2013) and the Nevado Cololo (Sanches, 2013; Veettil et al., 2016b) were also subjected extensive research. Glaciers in the Cordilleras Huaytapallana, Vilcabamba, Urubamba, Vilcanota, Carabaya (Peru), Cordillera Apolobamba (Peru-Bolivia), and Cordilleras Real, Tres Cruces and Nevado Santa Vera Cruz (Bolivia) belong to this group. In the southern wet outer tropics, air temperature oscillates more than 4^{o}C throughout the year and this region have lower humidity compared with the northern wet outer tropics, even though the mean annual temperature is the same as that in the northern wet tropics (about 1.6^{o}C, using CRU CL data expressed at 2 m above surface) (Sagredo and Lowell, 2012). The warm and wet season during the austral summer (November-April) differs about 1-2^{o}C from the cold and dry season (May-October) and precipitation occurs during the austral summer (particularly during December-March). Salzmann et al. (2013) observed a linear increasing trend in air temperature near the Cordillera Vilcanota since the 1950s and higher changes occurred in the case of maximum air temperature than minimum temperature (a slight negative trend in precipitation was also reported by the authors). The climate in this region is controlled by the seasonal oscillations of the ITCZ. The seasonality in precipitation becomes more distinguishable from the northern wet outer tropics towards the dry outer tropics through the southern wet outer tropics (Liu et al., 2013) and this difference in meteorological conditions influence mass balance variability along the Andean mountains. Accumulation and melting occurs during the wet season whereas sublimation is the main agent causing mass loss during the dry season (Favier et al., 2004a). However, Kinouchi et al. (2013) observed that mass loss by melting occurs during the dry season as well. Both sublimation and melting were observed to be higher during the transition between the dry and the wet seasons (Liu et al., 2013).

Lopez-Moreno et al. (2014) did a detailed work in the Cordillera Huaytapallana on the glacier retreat (six glaciers namely Huaytapallana, Chapico, Utcohuarco, Pitita, Marairazo and Azulcocha) and climate variations between 1984 and 2011 using Landsat images. They noticed a decrease of 56% of the initial glaciated area in the entire cordillera and observed the formation of numerous proglacial lakes nearby and the observed retreat was more than the double of the observed retreat in the Cordillera Blanca. The highest summit (Huaytapallana Glacier) lost only 42% of its area and glaciers having comparable elevation with smaller area retreated faster. Lopez-Moreno et al. (2014) also observed a significant increase in the maximum air temperature rather than changes in precipitation during El Niño years, similar to that observed near the Nevado Coropuna (Herreros et al., 2009), and the snowline has moved upwards by 93-157 m between 1984 and 2011. These values are much greater than reported from the northern wet outer tropics (Cordillera Huayhuash and Cordillera Raura) by McFadden et al. (2011), which indicates that the glaciers in the southern wet outer tropics are retreating faster than other parts of the outer tropics. Zubieta and Lagos (2010) predicted (simple linear trend analysis) that the glacial coverage in the Cordillera Huaytapallana would disappear completely by 2030, based on a study using Landsat images acquired between 1976 and 2006.

Glaciers in the entire Cordillera Vilcanota, the second largest mountain range in Peru, have lost an area of nearly 4 km^2/year during 1988-2010 and out of which the area of the Quelccaya Ice Cap, which is the largest tropical ice cap, declined at the rate of 0.6 km^2/year between 1980 and 2010 (31% of the initial area). The decline rate was more accelerated during 2000-2010 (Hanshaw and Bookhagen, 2014). Using aerial photographs and medium resolution satellite images, Salzmann et al. (2013) calculated a decrease of 25.5% of the glaciated area of the Quelccaya Ice Cap between 1962 and 2009 whereas an outlet glacier of Quelccaya, the Qori Kalis, lost 47% of the initial area during the same period. In the same paper, in contrary to higher precipitation sensitivity of the glaciers in the outer tropics as observed by Favier et al. (2004a, 2004b), the authors observed an enhanced temperature sensitivity than precipitation sensitivity on the ablation characteristics.

Recently, Herreros et al. (2009) and Veettil et al. (2016a) mentioned that the anomaly in precipitation during the strong El Niño in 1997/98 was not so higher but observed higher air temperature and the snowline of glaciers in the dry outer tropics increased (in the Cordillera Ampato, Peru). Salzmann et al. (2013) could not observe a considerable correlation between the ENSO indices and the area of glaciers in the Cordillera Vilcanota, as these glaciers might be highly influenced by the Amazon circulation than the Pacific circulation patterns. Hanshaw and Bookhagen (2014) observed the formation of numerous proglacial lakes in the Cordillera Vilcanota, particularly near the Nevado Ausangate, which is a clear sign of the continuous mass loss of glaciers in this region. The same study also observed that the anomalous decline in the area of glaciers (in 1998) might have related to ENSO in contrary to the observation by Salzmann et al. (2013).

Many glaciers in the Cordilleras Real and Apolobamba were studied recently using remote sensing (Soruco et al., 2009; Liu et al., 2013; Ribeiro et al., 2013; Sanches, 2013; Veettil et al., 2016b), ice core records (Angelis et al., 2003; Ramirez et al., 2003), and lichenometry (Rabatel et al., 2005a, 2006). The mean ELA of the glaciers in the Cordillera Real varies between 5200 to 5500 m a.s.l. for individual glaciers. Palaeoclimatic studies showed that an increased amount of precipitation from the Pacific played an important role in maintaining the ELA of the glaciers in this region during the late glacial period (Smith et al., 2011). Liu et al. (2013) calculated a decrease of 34.5% in the glacier-covered areas of the Cordillera Real (Huayana Potosi, Mururata, Charquini, Illimani and Serkhe Khollu glaciers) between 1987 and 2010 using Landsat, ALOS AVNIR-2 and ASTER GDEM data. Other glaciers such as the Chacaltaya have already disappeared by 2009 (Chevallier et al., 2011). The rate of retreat was, however, not uniform during this period. For example, higher glacier retreat was observed during 1987-1998 compared with 1999-2006 but an overall decreasing trend prevailed, possibly due to the rapid increase in temperature since the late 1970s. Smaller glaciers (Charquini Glacier, 2.4 km^2, for example) retreated faster compared with larger ones (Illimani Glacier, 20.8 km^2, for example). Perhaps a more complete record of glacier decline in the Cordillera Real during 1963-2006 was given by Soruco et al. (2009) based on photogrammetric measurements and they reported that the volume loss was 43% during 1963-2006, the surface area loss was 48% during 1975-2006, and the essential changes occurred during the late 1970s. Liu et al. (2013) pointed out that the glaciarized area on steeper slopes in the Cordillera Real retreated faster than those on gentle slopes and calculated the solar radiation dependency of mass balance. However, the orientation of slopes in the Cordillera Real with respect to solar radiation influences melting, which is true for all glaciers in the tropical Andes (south-facing slopes get higher levels of solar radiation during the melting season). The loss of glacier coverage in the Cordillera Real, as mentioned by

Ramirez et al. (2001) based on studies on Glaciar Chacaltaya, depends on the size of the glacier – the smaller ones will disappear first.

Veettil et al. (2016b) calculated the variations in annual snowline maximum, which is nearly the same as the ELA of glaciers in the outer tropics, of the Nevado Cololo and Nevado Huanacuni in the Cordillera Apolobamba between 1984 and 2012 and confirmed that despite the influence of cold La Niña events, the glaciers in this region are under decline. Similar trends were observed by Morizawa et al. (2013) in the case of Condoriri Glacier in the Cordillera Real. Jordan (1985) done the first photogrammetric height evaluation of glaciers in the eastern cordilleras of Bolivia and observed a small north to south gradient in snowline distribution in this region. Ribeiro and Simões (2010) studied seven glaciers in the Cordillera Tres Cruces (Jankho Loma, Jacha Pacuni, San Enrique, MallaChuma, Laramkkota, Campanani and Glacier c5250) using CBERS-2 images and concluded that the present day glacier distribution is about 245-352 m above the LIA distribution. Moreover, the same study also observed that the elevation of glaciers on the northeastern sides are about 100-270 m higher than glaciers on the south-western sides and the results from Veettil et al. (2016b)also supports these findings.

7.3. Dry outer tropics

The western cordilleras of Bolivia, southern Peru (south of 15^{o}S) and the north of Argentina and Chile (north of 18^{o}S) (glaciers in the cordilleras Huanzo, Chila, Ampato, Volcanica and Barroso) belong to this group and all glaciers in this region are situated above the mean annual 0^{o}C isotherm (Sagredo and Lowell, 2012). Strong sublimation occurs during dry periods and melting is one of the principal mass balance determining factor during the wet season. Some researchers consider the changes in the area of glaciers in the dry outer tropics as similar to that in the Cordillera Blanca (northern wet outer tropics) (Vuille et al., 2008a) and the mass balance of glaciers is highly sensitive to the seasonality and quantity of precipitation (Favier et al., 2004a; Sagredo et al., 2014). Based on δ^{18}O studies, Hardy et al. (2003) observed a strong relationship between precipitation and δ^{18}O, a weak relationship between temperature and δ^{18}O in the Sajama ice core and a strong correlation between ENSO and interannual precipitation variability in the region. Precipitation near the Cordillera Ampato depends on the tropical Atlantic circulation of air masses and the Pacific atmospheric circulation patterns also have a significant role in the climate variability in this region

(Herreros et al., 2009). Most of the precipitation (70-90%) occurs during the austral summer (December-March) and dry season occurs during the austral winter.

The Cordillera Occidental in southern Peru is having many ice-covered volcanoes and mountains in the Central Volcanic Zone (CVZ). Nevado Coropuna (6426 m a.s.l.) in the Cordillera Ampato is an ice-covered volcano which is studied extensively (Racoviteanu et al., 2007; Peduzzi et al., 2010; Ubeda 2011; Veettil et al., 2016a). Bromley et al. (2009, 2011), using lateral moraine deposits, studied the snowline variation during the late Pleistocene of the Nevado Coropuna and two neighbouring glaciated peaks namely the Solimana (6093 m a.s.l.) and the Firura (5498 m a.s.l.). Bromley et al. (2011) found that the snowline during the last glacial maximum (LGM) on the northern side was about 550 m lower than that of today and this difference on the southern side was about 640 m for Nevado Coropuna and similar trends were observed in the case of Solimana and Firura. Smith et al. (2005a) also give a complete regional synthesis of the snowlines of glaciated mountains in this region during the LGM. Ice-covered volcanoes in the Ampato-Sabancaya-Hualcahualca complex are having a complex glacier history due to the volcanic activities. Racoviteanu et al. (2007) calculated a decrease of 26% in the glacier area of Nevado Coropuna during 1962-2000 and Veettil et al. (2016a) calculated a decrease of 26.92% during 1986-2015. Recent variations in snowline patterns of Nevado Coropuna (Veettil et al., 2016a) were observed to be similar to that of Nevado Sajama (Veettil et al., 2016b).

The Cordillera Occidental in Bolivia is not an extensive mountain range as the Cordillera Oriental and the glaciers are limited to the Nevado Sajama (which is studied by many researchers) and the neighbouring ice-covered volcanoes (Parinacota, Pomerape, and the Volcan Guallatiri-Wallatiri complex) in the Bolivia-Chile border. Some researchers (Arnaud et al., 2001; Hardy et al., 2003; Veettil et al., 2016b) consider this region as a good study site to understand the ENSO influence on mountain glaciers in the tropical Andes, due to the particular distance from the Pacific Coast. Fluctuations in the annual snowline of Nevado Sajama seems to be very interesting; for example, the calculated snowline in 1986 October using Landsat image was 5460 m a.s.l. whereas in 1997 June the same was 5449 m a.s.l. (Arnaud et al., 2001). However, due to the strong El Niño during 1997/98 the snowline increased up to 5967 m a.s.l. in 1998 May. The same study observed, using aerial photographs, that the snowline was well below 5300 m a.s.l. between 1963 and 1972 (during the PDO negative regime) and then started rising (from 1984 onwards, after the PDO reversal in 1976). Veettil et al. (2016b) also calculated the variations in annual snowline of the Nevado Sajama between 1984 and 2012 and observed the variations in the snowline with the occurrence of ENSO during this period. Both studies (Arnaud et al., 2001; Veettil et al., 2016b) observed a general rise in the snowline during the respective study periods, with exceptional cases during the La Niña periods.

8. Conclusions

We have seen the glacier changes in the tropical Andes from the north to the south. Tropical Andean glacier records such as moraine chronology not only help us to understand the climate in the past but also provide an insight to what is happening in the present climate. Irrespective of various factors controlling glacier response to climate change, tropical glaciers provide complementary evidences on the magnitude of climate change. Other than information on long-term climate conditions, glacier records also provide insight to interannual and interdecadal variability such as ENSO and PDO in the past. Various methods applied so far, to understand the climate in the past from glacier records are lichenometry, ice core analysis, radiocarbon dating, thermoluminescence and Terrestrial Cosmogenic Nuclide analysis give us a coarse resolution picture of the evolution of glacier surfaces and glacier-climate interactions in the past.

In order to understand modern climate-glacier interactions, mass balance measurements and equilibrium line changes were widely used in the tropical Andes. ELA of tropical glaciers can be calculated using multiple parameters such as freezing height, precipitation, and temperature by applying various mathematical equations. The maximum snowline during dry season, which can be used as an approximate value of the equilibrium line, was also used for understanding mass balance fluctuations of the tropical Andean glaciers, particularly in the outer tropics (Rabatel et al., 2012; Veettil et al., 2016a, 2016b). Sublimation, which causes mass loss in a cold and dry climate, and the relative position of the $0^{o}C$ isotherm are major factors in determining the annual mass balance and cannot be ignored in the tropical Andes. A better understanding of the complex glacier-climate interactions in the tropical Andes can be achieved only by combining the three types of variables: morphological and topographical (altitude, latitude, slope, aspect, distance from the Pacific coast and the Amazon Basin), meteorological (precipitation, temperature, specific humidity, cloudiness, wind speed, sunshine duration, amount of solar radiation, radiation balance), and glaciological (glacier type, mass balance, snowline, ELA, albedo, sublimation, surface area, ice volume, ice thickness) and glacier response time can also vary with the above mentioned variables.

Both the inner and the outer tropics maintain homogenous temperature conditions throughout the year with a small seasonality (in the air temperature) in the outer tropics. Accumulation is limited to the precipitation season in the outer tropics whereas it occurs throughout the year in the inner tropics. There exist gradients in glacier response to climate change from the inner tropics to the outer tropics (north to south), from the Pacific coast to the Amazon Basin (west to east) and with elevation. Glaciers in the inner tropics are highly sensitive to temperature variations compared with those in the outer tropics and this explains the rapid disappearance of mountain glaciers in Colombia and Venezuela under global warming. In the outer tropics, on the other hand, glaciers are highly sensitive to precipitation variability than those in the inner tropics. Recent studies (Bradley et al., 2006; Pepin et al., 2015) observed an increased warming in the high elevation mountain regions (EDW) and despite the recent global warming hiatus, high elevation Andes continue to be getting warmed (Vuille et al., 2015). The inclusion of elevation as a variable while interpreting the cryosphere-climate interactions along the tropical Andes is important in this context.

Our observations on the gradients in glacier response to climate change in the tropical Andes fit well into the new framework for South American glacier classification by Sagredo et al. (2014). They used SEMB model which considers both surface energy balance and mass balance to calculate the sensitivity of equilibrium line towards variability in precipitation and temperature and has the advantage that gridded datasets such as the NCEP-NCAR reanalysis data can be used. However, this statistical model (Sagredo et al., 2014) has problems due to small number of glacier samples taken, coarse resolution and sparse coverage of meteorological datasets, and the constants used. These flaws can be overcome by using better glacier inventories (such as the RGI) and improve the resolution and coverage of meteorological datasets in the high mountain environments, such as from spaceborne sensors.

Glaciers in both the inner and the outer tropics undergo fluctuations in snowline during the phase changes of ENSO. However, a quantitative approach to estimate the differences in mass balance sensitivity of tropical Andean glaciers towards ENSO and other decadal variability from the north to the south is still missing. Estimation of snowline variations using remote sensing and calculating the correlation between snowline fluctuations and time series meteorological data is one approach to solve this problem. However, extensive cloud cover, particularly near the ITCZ, may influence the accuracy of the calculated snowline using optical satellite images such as Landsat and ASTER. Active remote sensing (radar, for example) which can penetrate thick cloud cover can be used to overcome these difficulties. It is observed recently that if El Niño occurs during the warm phase of PDO, the rate of increase in snowline elevation is high and this is true for both the inner and the outer tropics.

Another observation, that we have done so far, is that glaciers near and above the January ITCZ are retreating at a faster rate. Continuous retreat of glaciers in Venezuela and Colombia in the inner tropics and in the Cordillera Real (Bolivia) and the Cordillera Apolobamba (Bolivia-Peru) are the examples supporting this observation. Size and elevation are two important factors that control this retreat – glaciers with smaller size at lower altitudes would disappear first. This explains why the high altitude glaciers in Ecuador are retreating at a lower rate compared with those glaciers in Colombia and Venezuela. Furthermore, glaciers in the Cordillera Blanca and dry outer tropics show relatively lower rate of increase in the snowlines. It can be argued that the occurrence of La Niña events during the negative PDO retards snowline rise of glaciers near the Pacific coast compared with those situated towards the Amazon Basin in the east. However, further investigations are needed for a conclusion on these aspects.

Acknowledgements

This study was funded by Fundação de Amparo à Pesquisas do Estado do Rio Grande do Sul (FAPERGS).

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Table 1: Summary of studies mentioned in this study and the methodologies used on past glacier changes in the tropical Andes. The list is by country from north to south

Country	Study region	Authors	Method used	Comments/observations
Venezuela	Cordillera de Mérida	Polissar et al., 2006	Radiocarbon dating of lake sediments	1,500-year glaciation history in the Venezuelan Andes and LIA climate conditions
	Sierra Nevada de Mérida	Schubert and Clapperton, 1990	Radiocarbon dating	Quaternary glaciation studies in Colombia, Venezuela and Ecuador
	Central Venezuelan Andes	Schubert, 1974	Radiocarbon dating of moraine deposits	Dealing with glacial advances since the Late Pleistocene in the Venezuelan Andes.
	Cordillera de Mérida	Stansell et al., 2005	Radiocarbon dating	Dating sediment cores from lakes to understand late Pleistocene and Holocene glacial activity
	Cordillera de Mérida	Stansell et al., 2007	Accumulation-area ratio (AAR), Area-altitude-balance ratio (AABR)	Paleo-ELA and paleo-temperature reconstruction
	Humboldt Massif	Mahaney et al., 2000a	Stratigraphy of moraine deposits	Last Glacial Maximum ELA and subsequent retreat
	Timotes Andes, Mérida	Schubert and Vaz, 1987	Thermoluminescence	Pre-Mérida glaciation
	Cordillera de Mérida	Bezada, 1990	Thermoluminescence	The Late Pleistocene glaciation using thermoluminescence
	Pueblo Llano, Central Mérida Andes	Mahaney et al., 2000b	Optically-simulated Luminescence	Glacier chronology during the LGM using infrared simulated luminescence techniques
	Cordillera de Mérida	Angel et al., 2013	Terrestrial Cosmogenic Nuclide (^{10}Be)	Reconstruction of Pleistocene glacier changes
	Gavidia Valley	Angel et al., 2016	Terrestrial Cosmogenic Nuclide (^{10}Be)	Deglaciation during the LGM and the Oldest Dryas Stadial
	Sierra Nevada de Mérida	Carcaillet et al., 2013	Terrestrial Cosmogenic Nuclide (^{10}Be)	The first chronology of post-LGM glacier retreat in Venezuela
	Southeastern flank of the Mérida	Guzmán et al., 2013	Terrestrial Cosmogenic Nuclide (^{10}Be)	Dating of river terraces of Santo Domingo river and restoration of incision rate
	Venezuelan Andes	Wesnousky et al., 2012	Terrestrial Cosmogenic Nuclide (^{10}Be)	Quantification of the age of glaciation in the Venezuelan Andes
Colombia	Santa Marta, Cocuy and Nevado Ruiz-Tolima	Schubert and Clapperton, 1990	Radiocarbon dating	Quaternary glaciation studies in Colombia, Venezuela and Ecuador
	Ritacuba Negro, Sierra Nevada del Cocuy	Jomelli et al., 2014	Terrestrial Cosmogenic Nuclide (^{10}Be and ^{3}He)	Argued about a major glacier advance in the tropical Andes during the Antarctic cold reversal
Ecuador	Chimborazo area, Cordillera Occidental	Clapperton, 1985	Radiocarbon dating	Late quaternary moraine dating near the Chimborazo-Carihuairazo massif
	Chimborazo (Cordillera Occidental) and Antizana (Cordillera oriental)	Jomelli et al., 2009	Lichenometry	Maximum Glacial Extent in Ecuador
	Chimborazo ice-covered	Schotterer et al., 2003	Ice core records	Glacier records to estimate extreme climate conditions such as

	volcano, Cordillera Occidental			volcano eruptions and draught
	Chimborazo ice-covered volcano, Cordillera Occidental	Vuille et al., 2003b	Ice core records	Precipitation and temperature records and ENSO signals from ice core records based on $\delta^{18}O$
Peru	Cordillera Blanca	Seltzer, 1987	Radiocarbon dating	Moraine dating in the Cordillera Blanca and estimating climate change
	Cordillera Blanca	Rodbell, 1992	Lichenometry and Radiocarbon dating	Holocene glacial chronology of the Cordillera Blanca. Observed the Holocene and LIA advances
	Cordillera Blanca	Solomina et al., 2007	Lichenometry and geomorphic studies	Dating LIA moraine on Pacific-facing and Atlantic-facing slopes and glacial advances
	Cordillera Blanca	Jomelli et al., 2008	Lichenometry	Chronology of glacier advances during the LIA
	Nevado Huascaran, Cordillera Blanca	Thompson et al., 1995	Ice core records	Isotopic records and Atlantic SST interpretation during the Last Glacial Stage
	Nevado Huascaran, Cordillera Blanca	Henderson et al., 1999	Ice core records	An attempt to estimate El Nino events, and climate variability over Amazonia and western tropical Atlantic from ice core records.
	Cordillera Vilcanota	MercerandPalacios, 1972	Radiocarbon dating	Last Glacial Maximum and LIA advances
	Huancané outlet glacier of Quelccaya Ice Cap, Cordillera Vilcanota	Malone et al., 2015	Numerical models	Quantify glacier length response to climate change during LIA and Younger Dryas (YD)
	Quelccaya Ice Cap, Cordillera Vilcanota	Thompson et al., 1984	Ice core records	ENSO events recorded in the Quelccaya ice core, method can be used for at least 1500 years
		Thompson et al., 1985	Ice core records	Interpreting 1500-year tropical precipitation record
		Thompson et al., 2006a	Ice core records	Interpreting the abrupt climate change in the tropical Andes since the last two millennia
	Nevado Coropuna, Cordillera Ampato	Thompson et al., 2006b	Ice core records	Interpreting the abrupt climate change in the tropical Andes
	Nevado Coropuna, Cordillera Ampato	Herreros et al., 2009	Ice core records	Ice core analysis showed the dependency of precipitation on easterly circulation from the tropical Atlantic and identified ENSO
	Nevado Coropuna, Nevado Solimana and Nevado Firura in the Cordillera Ampato	Bromley et al., 2011	Geomorphic-numeric approach	Snowline reconstruction during the Pleistocene
	Nevado Coropuna, Cordillera Ampato	Bromley et al., 2009	Cosmogenic ^{3}He analysis	Glacier advance and retreat during Last Glacial Maximum and Late-Glacial period
	Cordillera Blanca	Farber et al., 2005	Terrestrial Cosmogenic Nuclide (^{10}Be)	Calculated ages of Pleistocene glacial moraines
	Cordillera Huayhuash	Hall et al., 2009	Terrestrial Cosmogenic Nuclide (^{10}Be), ASTER images, aerial	Geochronology of Quaternary glaciations

			photographs and GPS	
Bolivia	Cordillera Apolobamba	Gouze et al., 1986	Radiocarbon dating	Estimating the maximum and minimum ages of late Quaternary moraines
	Nevado Sajama	Thompson et al., 1998	Ice core records	Estimating climate history of the tropics since the Last Glacial Stage from ice core records
	Nevado Sajama	Bradley et al., 2003	Ice core records	Pacific SST and hence ENSO variability recorded in ice cores
	Charquini glaciers, Cordillera Real	Rabatel et al., 2006	Lichenometry and AAR	Moraine dating and ELA reconstruction
	Charquini glaciers, Cordillera Real	Rabatel et al., 2005a	Lichenometry	Dating LIA glacier fluctuations
	Nevado Illimani, Cordillera Real	Hoffmann et al., 2003	Ice core records	Four ice core records from Huascaran, Quelccaya, Illimani and Sajama showed identical decadal variability in the 20th century
	Nevado Illimani, Cordillera Real	Knüsel et al., 2003, 2005	Ice core records	Dating and estimating 20th century ENSO signals respectively
	Nevado Illimani, Cordillera Real	Ramirez et al., 2003	Ice core records	Established strong resemblance with other ice core records in the tropical Andes
	Bolivian Altiplano	Quesada et al., 2015	^{18}O isotope analysis	How the ^{18}O content of the Andean glaciers are influenced by Bolivian palaeolake evaporation during the last Deglaciation
	Cerro Azanaques, Cordillera Central	Martin et al., 2015	Terrestrial Cosmogenic Nuclide (^{10}Be)	Cosmogenic ^{10}Be concentrations for estimating age Cerro Azanaques and Challapata fan-delta
	Bolivia-Peru	Smith et al., 2005	Terrestrial Cosmogenic Nuclide (^{10}Be)	Dating the Last Local Glacial Maximum (LLGM) in the Andes of Peru and Bolivia

Table 2: Details of satellite data/maps used in glacier mapping/mass balance determination studies in the topical Andes described in this review

Region in the tropical Andes	**Authors**	**Glacier/ Massif**	**Country**	**Satellite data/ Aerial photography and spatial resolution/scale**	**Quantification of cartographic error/vertical accuracy**	**Brief description**
Inner tropics	Schubert (1998)	Three massifs namely Pico Boliver, Pico La Concha and Picos Humboldt/Bonpland	Venezuela	1952 Aerial photography Landsat MSS - 60 m	Not given	Detailed glacier atlas of Venezuela
	Braun and Bezada (2013)	Humboldt Glacier	Venezuela	GPS measurements in 2011	0.019 km^2 in 2009 0.017 km^2 in 2011	Mapped the glacier area in 2011 and compare with the previous studies
	Ceballos et al. (2006)	Sierra Vevada de Santa Marta,Sierra Nevada del Cocuy, Nevado del Ruiz, Nevado de Santa Isabel, Nevado del Tolima and Nevado del Huila	Colombia	Aerial photography Landsat TM – 30 m Ground Penetrating radar (GPR) of 10 MHz	Not mentioned	Documented the shrinkage of six major ice masses in Colombia during 1959-2002 using aerial photography, Landsat images, GPR and irregularly spaced GPS measurements
	Morris et al. (2006)	Pico BonplandMassif (Siniguis) Ruiz-TolimaMassif, Sierra Nevada de Santa Marta andSierra Nevada del Cocuy	Venezuela and Colombia	Landsat MSS – 60 m Landsat TM – 30 m ASTER – 30 m	Not mentioned	Glacier retreat in Venezuela and Colombia using Landsat and ASTER images and compared the results from previous studies from 1950s to 2000-2003
	Poveda and Pineda (2009)	Sierra Nevada de Santa Marta, Sierra Nevada del Cocuy, Nevado del Ruiz, Nevado de Santa Isabel, Nevado del Tolima and Nevado del Huila	Colombia	Landsat TM, ETM+ - 30 m	Not mentioned	Retreat of mountain glaciers and ice caps in Colombia during 1989-2005 and predicted that these glaciers will disappear by 2020
	Jordan et al. (2005)	Cotopaxi Volcano, Cordillera Oriental	Ecuador	Aerialphotographs (1:30 000)	Not mentioned	Glacier retreat of Cotopaxi Ice covered volcano between 1956 and 1976 using photogrammetry
	Veettil (2012)	Chimborazo, Cordillera Occidental	Ecuador	Landsat TM – 30m Landsat ETM+ - 30m	Not mentioned	Changes in glacial aerial extent with ENSO phase changes

	Veettil et al. (2014a)	Cotopaxi Volcano, Cordillera Oriental	Ecuador		Not mentioned	Calculating the combined influence of ENSO and PDO on the annual snowline of the Cotopaxi ice-covered volcano
Northern wet outer tropics	Georges, 2004	Cordillera Blanca	Peru	SPOT XS – 20 m 50-m contour lines		20th century glacier fluctuations in the Cordillera Blanca
	Silverio and Jaquet (2005)	Cordillera Blanca	Peru	Landsat TM – 30 m	Nearly 10%	Glacial cover mapping using NDSI images by applying a suitable threshold between 1987 and 1996
	Veettil et al., 2016e	Southern Cordillera Blanca	Peru	Landsat TM, ETM+ and OLI – 30 m IRS LISS III – 23.5 m ASTER GDEM	Vertical accuracy of DEM – 20 m	Annual snowline variations in Southern Cordillera Blanca during 1984-2015
Southern wet outer tropics	Ribeiro et al. (2013)	Nevado Illimani, Cordillera Real	Bolivia	ALOS PRISM – 2.5 m ALOS PRISM DEM – 10 m Aerialphotograph (1:29 200) Aerialphotograph (1:39 500)	RMSE 1.25 m Vertical accuracy 6.14 m RMSE 1.67 m RMSE 0.62 m	Calculated the ice extent of Nevado Illimani during 1963 and 2009 using photogrammetry
	Hanshaw and Bookhagen (2014)	Quelccaya Ice Cap, Cordillera Vilcanota	Peru	Landsat MSS – 60 m Landsat TM – 30 m Landsat ETM+ - 30 m ASTER – 15 m Corona KH-9 – 6-9 m ASTER GDEM – 30 m	<5% for glaciers >1 km^2; Overall error of 2-6% using automated methods	Calculated area, snowline and glacial lake area during 1975-2012
	Veettil et al. (2016b)	Nevado Cololo, Cordillera Apolobamba	Bolivia	Landsat TM – 30m Landsat ETM+ - 30m IRS LISS III – 30 m ASTER GDEM - 30 m	Vertical accuracy of DEM – 20 m	Variations in annual snowline and correlates with the combined influence on ENSO and PDO
	Peduzzi et al., 2010	Nevado Coropuna, Cordillera Ampato	Peru	Landsat MSS – 60 m Landsat TM – 30 m ASTER – 30 m	Error margin varied with datasets used	Ice thickness, area, and volume changes during 1955-2008
				Landsat TM – 30m		Annual changes in area and dry

Dry outer tropics	Veettil et al. (2016a)	Nevado Coropuna, Cordillera Ampato	Peru	Landsat ETM+ - 30m IRS LISS III - 23.5 m ASTER GDEM – 30 m	Vertical accuracy of DEM – 20 m	snowline during 1986-2014
	Veettil et al. (2016b)	Nevado Sajama, Cordillera Occidental	Bolivia	Landsat TM – 30m Landsat ETM+ - 30m IRS LISS III – 30 M ASTER GDEM	Vertical accuracy of DEM – 20 m	Variations in annual snowline and correlates with the combined influence on ENSO and PDO

Table 3: Glacier nomenclature in three massifs in Venezuela given by Jahn (1925) and later modified by Schubert (1992, 1998) (in brackets)

Pico Bolivar Massif	Pico La Concha Massif	Pico Humboldt/Bonpland Massif
Timoncito Glacier(Timoncito-Hermanas Glacier)	Garza Glacier (Ño León Glacier)	Plazuela Glacier (Nuestra Señora Glacier)
Karsten Glacier (El Encierro Glacier)	Mucuy Glacier (Coromoto Glacier west remnant)	Sievers Glacier (Sinigüis Glacier)
Hermanas Glacier		Codazzi Glacier
Espejo Glacier		Laguna Verde Glacier (Coromoto Glacier east remnant** orHumboldt Glacier***)
Bourgoin Glacier*		

* Together with Karsten Glacier later named as El Encierro Glacier

** Name given by Schubert (1992,1998)

*** Local name and is the only remaining glacier in Venezuela

CAPITULO III

Metodologia

1. ÁREA DE ESTUDO

Nesta pesquisa, foram consideradas para estudo cinco geleiras dos Andes tropicais, sendo duas delas situadas nos trópicos externos secos e outras três distribuídas entre os trópicos internos, trópicos externos úmidos do norte e trópicos externos úmidos do sul. Os Andes são considerados como uma linha divisória entre o Oceano Pacífico e a Bacia do Amazonas, com uma orientação norte-sul (Laraque *et al.*, 2007). Devido à localização geográfica, teoricamente, as geleiras da Cordilheira ocidental são mais influenciadas pela circulação atmosférica do Pacífico, enquanto as geleiras da cordilheira oriental são mais influenciadas pela circulação da Amazônia e do Atlântico. Além disso, há também a hipótese de que as geleiras nos trópicos internos se comportem com diferentes taxas de ablação em comparação com aquelas geleiras nos trópicos externos. As subseções a seguir descrevem a localização geográfica e as condições climáticas dos locais de estudo considerados para esta pesquisa. A **figura 1** mostra a localização das geleiras selecionadas.

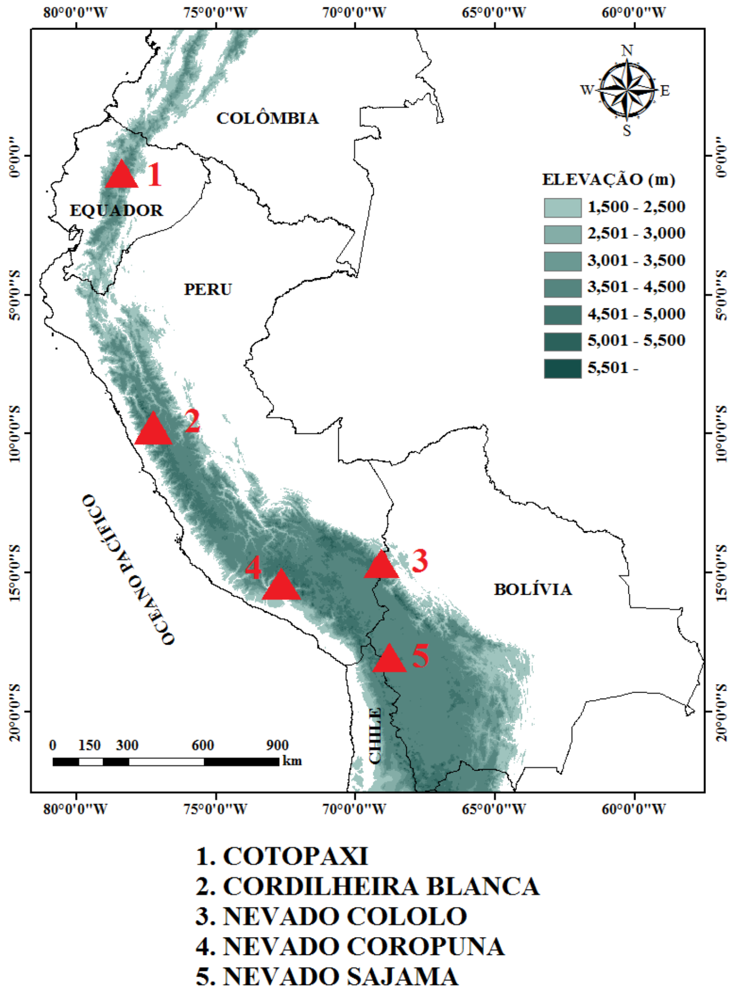

Figura 1—As localizações das geleiras selecionadas

1.1. Local de estudo 1 - Trópicos internos: Vulcão (Nevado) Cotopaxi, Equador

No Equador, as geleiras estão situadas nos trópicos internos em duas cordilheiras - Cordilheira Ocidental (Lat: 0º22' N - 1º29' S; Lon: 78º20' W - 78º48' W) e Cordilheira Oriental (Lat: 0º1' N - 2º20' S; Long: 77º54' W - 78º33' W). O Cotopaxi na Cordilheira Oriental, localizado a cerca de 60 km a sudeste da capital Quito, é um dos maiores vulcões do mundo e é o segundo pico mais alto do Equador (5897 m a.n.m; 0º40' S, 78º25' O). A **figura 2** mostra a localização do Cotopaxi e a distribuição da altitude no Equador. É um vulcão coberto de gelo e está amplamente documentado no Equador e suas flutuações na linha de neve têm sido estudadas durante os últimos 30 anos. A Cordilheira Oriental está diretamente exposta aos ventos úmidos do leste da bacia do Amazonas. Cerca de 20 geleiras irradiam para fora da sua calota de gelo em todas as direções (Jordan *et al*., 2005). Entre 1976 e 1997, o Cotopaxi havia perdido cerca de 30% de sua área de geleira (Jordan *et al*., 2005). Erupções mais recentes foram relatadas em 1742-1744, 1768, 1877 (Jordan *et al*., 2005) e 2015. Cáceres *et al*. (2004) tentou quantificar a descarga de gelo que pode ocorrer no futuro devido ao fenômeno laharitico causado por erupções vulcânicas. A imagem de satélite do Cotopaxi e suas respectivas regiões a serem estudadas podem ser vistas na **figura 3**. Cerca de 4% das geleiras tropicais do mundo estão situadas nos Andes equatorianos (Kaser, 1999). O vulcão Antizana situado na cordilheira oriental passou por um recuo maciço entre os anos de 1956 e 1998 (Francou *et al*., 2000). A geleira Antisana 15 avançou entre 1999 e 2001 devido às condições extensas de La Niña que prevaleciam na região (Francou *et al*., 2000, 2004). Esta pesquisa tentará encontrar uma correlação entre a ocorrência de ENOS (e ODP) e alterações de linha de neve na calota de gelo do vulcão Cotopaxi com base em sensoriamento remoto e conjuntos de dados meteorológicos.

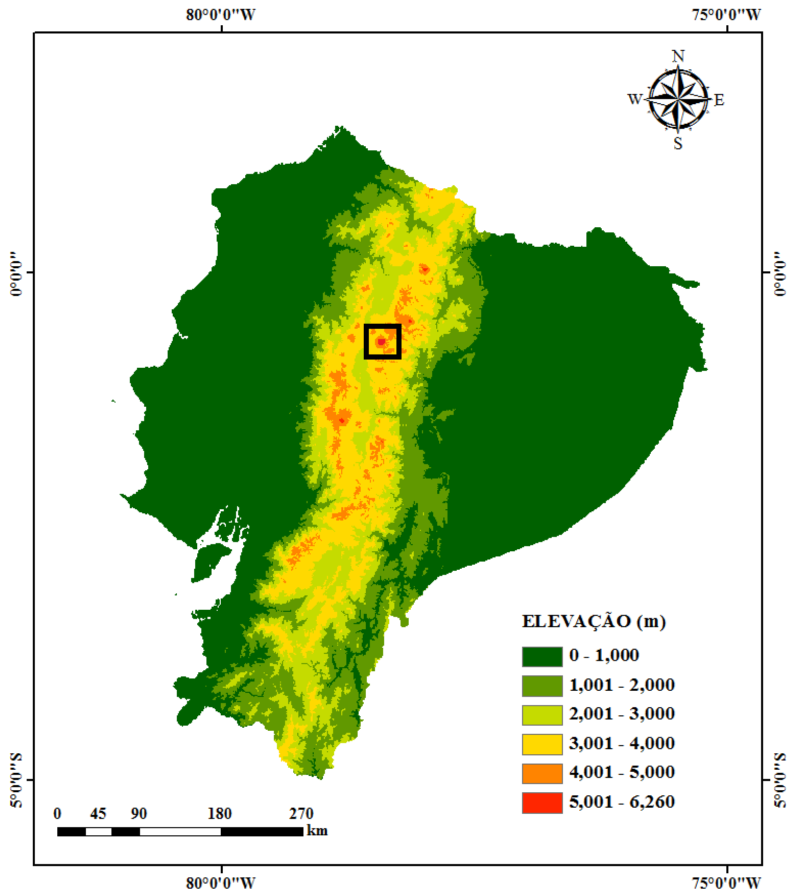

Figura 2—A localização do Cotopaxi e a distribuição da altitude no Equador

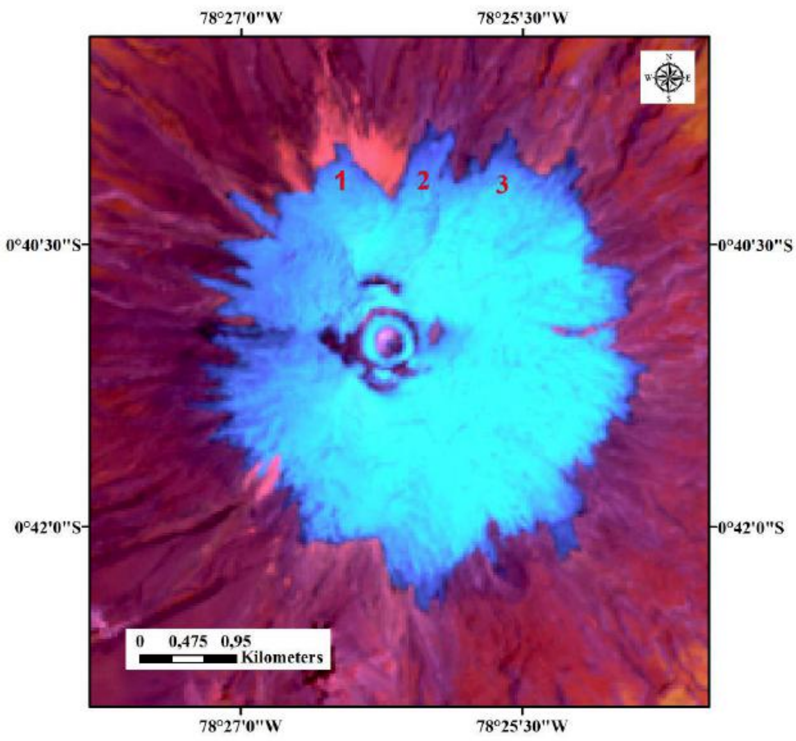

Figura 3—A imagem de satélite de Cotopaxi e suas respectivas regiões a serem estudadas

Nesta região existem dois máximos de precipitação por ano: sendo o principal nos meses de março, abril e maio, e o secundário nos meses de setembro, outubro e novembro. As variações sazonais nas temperaturas não são significativas, mas a variabilidade interanual é consideravelmente grande. O vento é o principal fator de sazonalidade na cordilheira oriental do Equador (Francou *et al*., 2004), também aqui, a precipitação é alterada pelo sistema da montanha andina, pelo fenômeno de mesoescala dos ventos do vale e pelo fenômeno das correntes oceânicas (ENOS e Corrente de Humboldt) (Bendix e Lauer, 1992), que por conseqüência afeta as mudanças de balanço de massa. A temperatura do ar é outro fator que controla o equilíbrio de massa-energia na região e, portanto, é mais sensível às variações climáticas do que nos trópicos externos (Favier *et al*., 2004a, 2004b; Jomelli *et al*., 2009). A incidência de radiação de ondas curtas é máxima durante os períodos próximos ao equinócio (março-abril e setembro) e a ausência de precipitações sólidas durante este período causará uma taxa de fusão significativa (Rabatel *et al*., 2013).

1.2. Local de estudo 2 - Trópicos externos úmidos do norte: Nevados Caullaraju-Pastoruri, Cordilheira Branca, Peru

Como parte da América do Sul, a Cordilheira Branca nos Andes estende-se por cerca de 180 km de comprimento com 30 km de largura, entre as latitudes 8°30´ - 10°10´S e longitudes 77º00'-78º00'O. Ela está localizada no Estado peruano de Ancash, 400 quilômetros ao norte da capital Lima, conta com 27 picos que atingem altitudes superiores a 6.000 metros e mais de 200 picos que excedem 5.000 m. O conjunto de geleiras Caullaraju-Pastoruri Glaciar (Lat: 9º55'45" S; Long: 77º12'18" O) no sul da Cordilheira Branca, parte dos Andes peruanos, é considerado nesta pesquisa (**figuras 4 e 5**). Avalanches catastróficas recentes ocorreram em 1962 e 1970 nesta região, onde o clima é típico dos trópicos externos. A Cordilheira Branca é caracterizada por pequenas variações sazonais em sua temperatura, porém com grandes variações diárias neste parâmetro, bem como pela alternância entre uma estação seca (maio a setembro) e uma estação chuvosa (outubro a abril). A estação chuvosa é responsável por 70-80% da precipitação anual (Kaser e Georges, 1997).

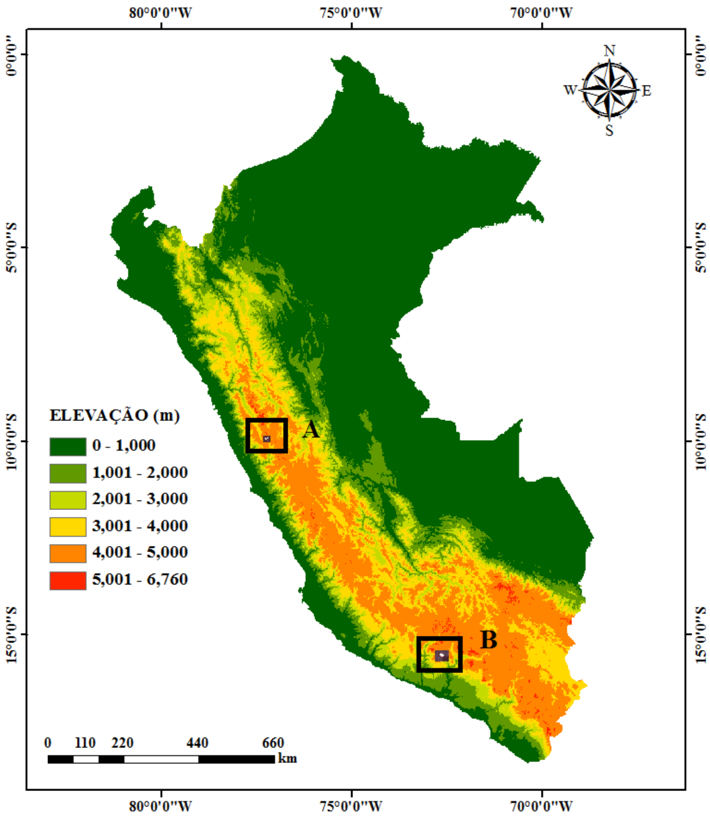

Figura 4—Localização dos Nevados Caullaraju-Pastoruri (A) e Nevado Coropuna (B) e a distribuição da altitude no Peru

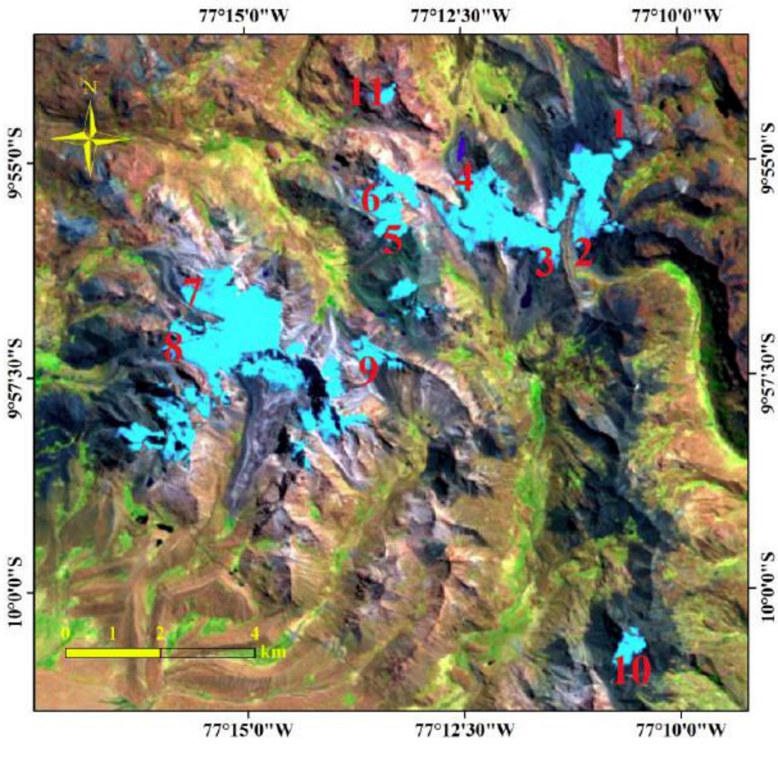

Figura 5—A imagem de satélite dos Nevados Caullaraju-Pastoruri e suas respectivas regiões a serem estudadas

Apesar da proximidade dos nevados com o Oceano Pacífico, a persistência de ventos do leste determina a fonte de umidade para a precipitação Andina e é predominantemente derivada do Atlântico (Johnson, 1976), sendo o clima do Peru altamente influenciado pelas montanhas andinas. Considerando que o ENOS é um fenômeno principalmente controlado pelas temperaturas da superfície do mar na bacia do Pacífico, este fenômeno muitas vezes reflete os eventos de aquecimento/arrefecimento no Atlântico tropical com defasagem de seis a oito meses. As anomalias de TSM no Atlântico tropical são substancialmente mais fracas do que as observadas no Pacífico equatorial em associação como ENOS. O acúmulo de massa ocorre apenas durante a estação chuvosa, predominantemente nas partes superiores das geleiras, enquanto a ablação ocorre durante todo o ano. As geleiras situadas nos trópicos externos e subtrópicos são consideradas como insensíveis à temperatura. O leste do Oceano Atlântico e a Bacia Amazônica são as duas principais fontes de precipitação nos Andes tropicais, principalmente devido os ventos sazonais do leste (Vuille e Keimig, 2004). Há 15 estações meteorológicas operadas pelo SENAMHI peruano (*Servicio Nacional de Meteorologia e Hidrologia*) distantes 60 quilômetros do local de estudo. As variações sazonais de temperatura são pequenas enquanto as variações de precipitação são maiores, sendo que cerca de 70-90% da precipitação ocorre durante o verão austral (dezembro-março), porém a estação seca nos Andes tropicais do Peru ocorre durante o inverno austral. As taxas de precipitação mais elevadas ocorrem com mais freqüência nas encostas viradas para o leste, provavelmente devido ao maior transporte de umidade da bacia Amazônica. As diminuições das taxas de precipitação foram observadas durante os fortes eventos de El Niño em 1982-1983 e 1992 enquanto o forte El Niño de 1997 não interferiu nas taxas de precipitação observadas (Herreros *et al.*, 2009). Infelizmente, a maioria das estações meteorológicas e hidrológicas parou de funcionar ou tem dados incompletos.

1.3. Local de estudo 3 - Trópicos externos úmidos do sul: Nevado Cololo, Cordilheira Apolobamba, Bolívia

Na Bolívia, considera-se que as geleiras estão situadas nos trópicos externos por apresentarem características como pequena variabilidade da temperatura, influxo de alta radiação solar durante todo o ano e sazonalidade de umidade e precipitação (Rabatel *et al.*, 2012). O Nevado Cololo, (14º50´S, 69º06´O; 5859 m a.n.m), na Cordilheira Apolobamba, é outro local de estudo, utilizado para entender a resposta das geleiras para flutuações climáticas nos trópicos externos úmidos do sul, caso ocorram (**figuras 6 e 7**). Sob o ponto de vista glaciológico, as mudanças de ablação podem ser divididos em três tipos (Rabatel *et al*., 2012), sendo: (1) as taxas de fusão mais alta, devido à radiação solar (outubro-dezembro), (2) a taxa de ablação maior devido ao degelo (janeiro-abril) e (3) as taxas de ablação limitadas devido à perda de energia por radiação de onda longa (maio-agosto). Nota-se que se a precipitação de neve ocorrer de maio a agosto, podendo permanecer durante a estação seca (Rabatel *et al*., 2012), isto acrescenta dificuldades em mapear o término da geleira usando sensoriamento remoto.

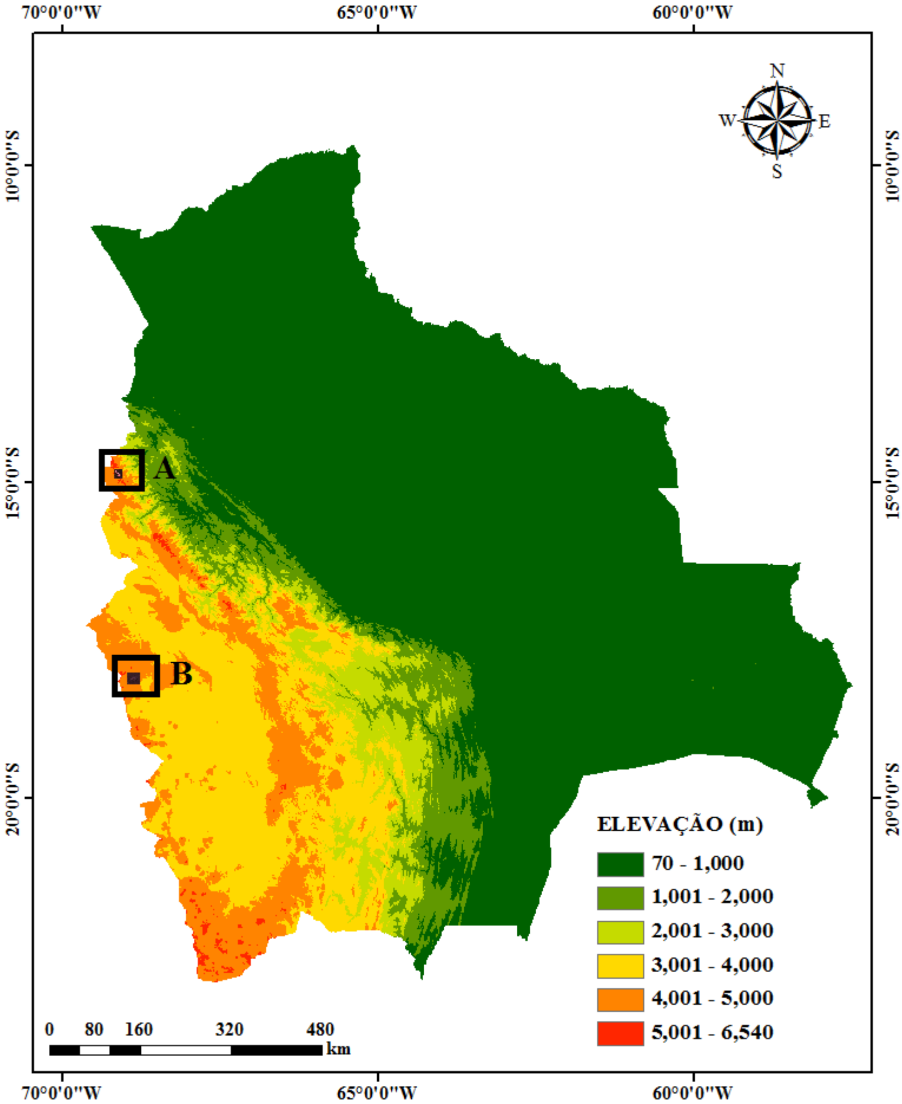

Figura 6—Localização do Nevado Cololo (A) e Nevado Sajama (B) e a distribuição da altitude na Bolívia

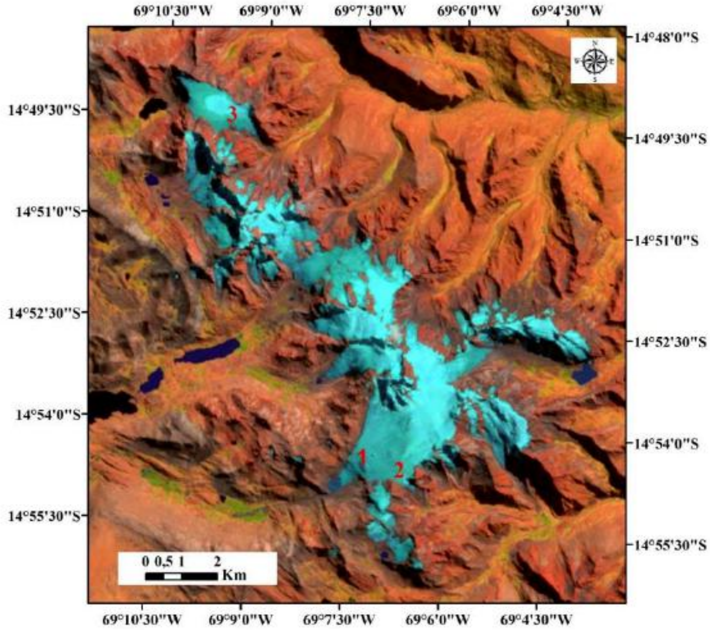

Figura 7—A imagem de satélite de Nevado Cololo e suas respectivas regiões a serem estudadas

O clima na Bolívia varia do tropical ao deserto frio, dependendo da altitude (Seiler *et al.*, 2013a), a temperatura média anual varia de 0º a 30ºC e a precipitação anual varia de menos de 300 mm a 3.000 mm ocorrendo principalmente nos meses de dezembro a março. A precipitação na Bolívia e sua variabilidade interanual estão ligadas às anomalias de TSM tropicais e às circulações atmosféricas (Arnaud *et al.*, 2001; Vuille, 1999). O verão austral (DJF) é caracterizado por um sistema de baixa pressão que aumenta os ventos (alísios) do leste e transportam a umidade do Atlântico (trópicos do norte) para o continente. Este teor de umidade é desviado pelos Andes e transportado para o sul causando precipitação reforçada no Oceano Atlântico (Marengo *et al.*, 2004). Com a liberação de calor de condensação sobre a Amazônia ocorre simultaneamente uma formação de um sistema de alta pressão boliviana de nível superior causado pelas encostas andinas, este processo faz com que ocorra o transporte avançado de umidade da Amazônia para as terras altas e terras baixas da Bolívia (Seiler *et al.*, 2013a; Vuille, 1999). No inverno austral (JJA), há menos transporte de umidade e ocorre a partir do norte do Atlântico tropical para o continente e as frentes frias penetram nas terras baixas bolivianas diminuindo assim a temperatura e a precipitação limitada (Garreaud, 2009). Os ventos de oeste que prevalecem na Bolívia impedem o transporte de umidade para a Cordilheira dos Andes durante o inverno austral (Vuille, 1999). As três principais fontes de variabilidade climática na Bolívia são: (1) Oscilação Decenal do Pacífico (ODP), (2) El Niño-

Oscilação Sul (ENOS) e (3) Oscilação Antártica (OA) (Seiler *et al*., 2013a). Com base em observações meteorológicas, Seiler *et al*. (2013b) afirmaram que o clima boliviano está aquecendo a uma taxa de 0,1ºC a cada década e segue os padrões ODP. Se há tal aquecimento, os indicadores de clima - geleiras de montanha – devem sofrer alterações de balanço de massa, e isto é uma das propostas a serem analisadas nesta pesquisa.

1.4. Local de estudo 4 - Trópicos externos secos: Nevado Coropuna, Cordilheira Ampato, Peru e Nevado Sajama, Bolívia

O Nevado Coropuna (Lat: 15 ° 24'-15 ° 51 'S; Long: 71 ° 51'-73 ° 00' O) na Zona Vulcânica Central (ZVC) na Cordilheira Ampato, sul do Peru, é considerado nesta pesquisa (**Figura 8**). A Cordilheira Ampato é composta de 93 geleiras, com uma espessura média de cerca de 40 m e uma área de superfície total de 146,73 km^2 com base em fotografias aéreas de 1962. O Nevado Coropuna é o pico mais alto (6426 m a.n.m) na Cordilheira Ampato e o vulcão mais alto no Peru (Racoviteanu *et al*., 2007). Muitas pessoas na parte norte da cidade de Arequipa dependem do derretimento das geleiras para o seu abastecimento de água potável, esta água é proveniente da ablação das geleiras do Nevado Coropuna. O recente encolhimento das geleiras na Cordilheira dos Andes peruanos iniciou no segundo semestre de 1980 (Huh *et al*., 2012; Salzmann *et al*., 2013). Racoviteanu *et al*. (2007) descobriram que o tamanho do Coropuna diminuiu de 82,6 km^2 em 1962 para 60,8 km^2 em 2000. O balanço de massa das geleiras nessa região depende muito das variações na precipitação (Wagnon *et al*., 1999).

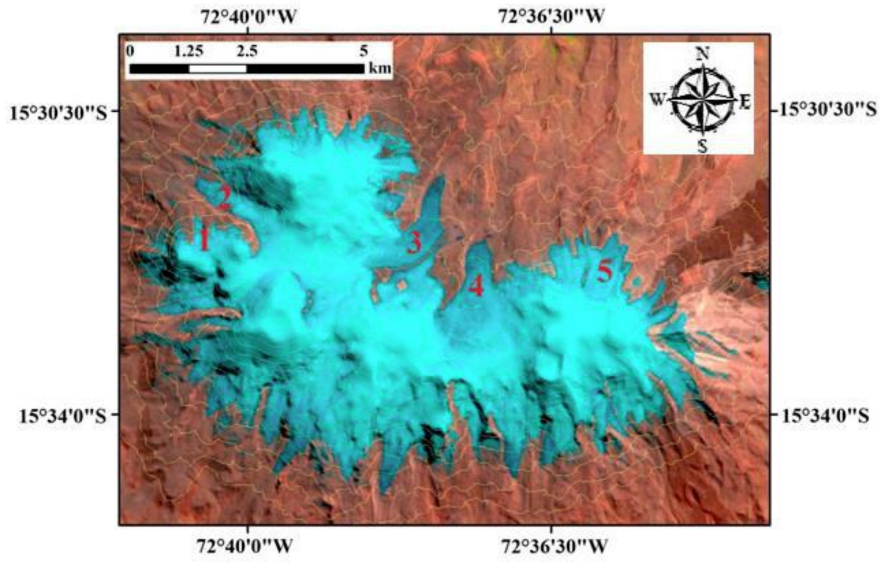

Figura 8—A imagem de satélite de Nevado Coropuna e suas respectivas regiões a serem estudadas

A precipitação na região Coropuna depende principalmente da circulação de massas de ar vindas do leste do Oceano Atlântico tropical (Herreros *et al*., 2009). No entanto, no Pacífico, os padrões de circulação atmosférica são significativos na determinação do clima nesta região. Como outras geleiras em regiões subtropicais e tropicais exteriores, Nevado Coropuna também está tendo uma ALE acima do 0°C isotérmica. Geleiras situadas nos trópicos externos e subtrópicos são, portanto, consideradas como temperatura insensível (Kaser, 1999). O leste do Oceano Atlântico e da bacia amazônica controla a precipitação nos Andes tropical, principalmente pelos ventos sazonais do leste (Vuille e Keimig, 2004). Para o monitoramento da região, há 15 estações meteorológicas operadas pelo serviço meteorológico e hidrológico nacional peruano ao longo de 60 km do local do estudo. A variação sazonal da temperatura é pequena, porém a precipitação é maior e chega a cerca de 70-90% da precipitação para o período de verão austral (dezembro-março). O inverno austral, que é a estação mais seca nos Andes tropicais do Peru, ocorre durante os meses maio-agosto. As taxas de precipitação mais elevadas foram observados nas encostas viradas para o leste, provavelmente devido à maior transporte de umidade da Bacia amazônica. A diminuição nas taxas de precipitação foi observada durante os eventos de El Niño forte durante 1982-1983 e 1992, enquanto no forte El Niño em 1997 não se observou interferência nas taxas de precipitação (Herreros *et al*., 2009). Infelizmente, muitas das estações meteorológicas e hidrológicas da região pararam de funcionar ou tem conjunto de dados incompletos.

O vulcão Sajama (18º06'S, 68º50'O, 6542 m de altitude), na região de Oruro, província de Sajama, na Bolívia é um estratovulcão na Zona Vulcânica Central (ZVC) no Andes Central e é o ponto mais alto da Bolívia (**figura 9**). A Sajama é a calota de gelo mais meridional na zona intertropical (Arnaud *et al*., 2001). Nesta região, o clima predominante é o semiárido e a precipitação anual é de cerca de 350 mm (Arnaud *et al*., 2001). A ablação e a acumulação atingem o seu ponto máximo durante o período chuvoso (outubro a março) (Ribstein *et al*., 1995). Os avanços máximos das geleiras ocorreram durante a segunda metade do século XVII (Rabatel *et al*., 2006) e acarretaram grandes recessões durante o século XX, mais intensivamente após a década de 1940. Localizadas a cerca de 100 quilômetros ao leste da costa peruana do Oceano Pacífico, no planalto de Altiplano, as calotas de gelo da montanha acima mencionadas são excelentes para estudar os impactos do ENOS (Arnaud *et al*., 2001).

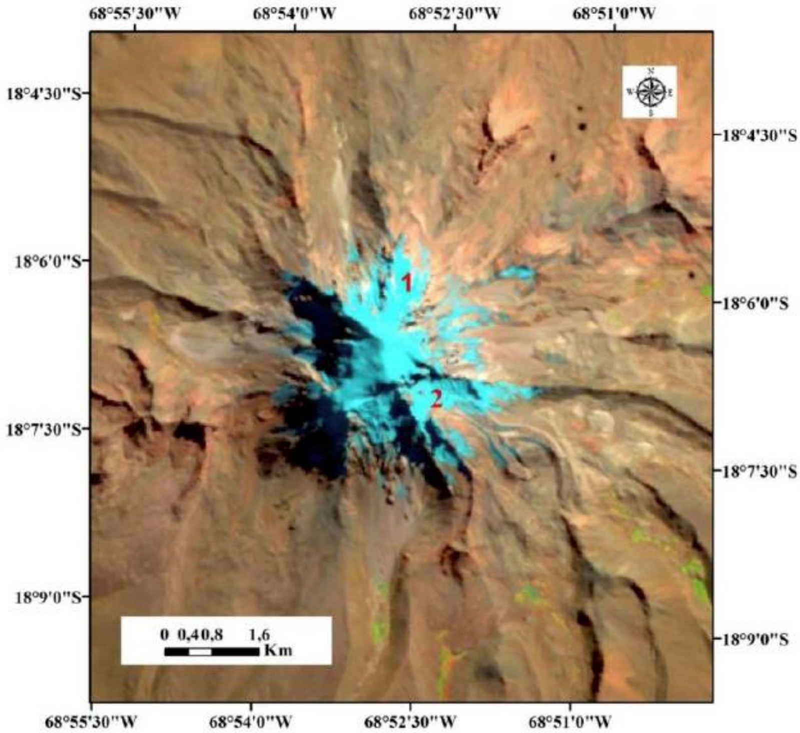

Figura 9—A imagem de satélite de Nevado Sajama e suas respectivas regiões a serem estudadas

A taxa de precipitação média mensal tem alta gradiente longitudinal de oeste para leste e, portanto, a precipitação observada é muito diferente nas duas cordilheiras da Bolívia. A precipitação ocorre pelo mecanismo de Monção amazônica que contribui com cerca de 80% da precipitação anual de outubro a abril, em particular na Cordilheira Real (Rabatel *et al*., 2006). A taxa de precipitação superior que ocorre perto da Cordilheira Oriental se deve ao teor de umidade reforçada transportada pela circulação atmosférica da Bacia Amazônica e parece estar enfraquecendo para o Nevado Sajama.

2. DADOS

Os dados usados são de dois tipos - de sensoriamento remoto e de meteorologia. Os dados de sensoriamento remoto incluem as imagens multiespectrais de vários sensores no período desde 1974 e modelos digitais de elevação (MDE). Os dados meteorológicos incluem precipitação e temperatura do ar durante o período de estudo. Os índices de ENOS e ODP também estão incluídos nos dados meteorológicos. Os detalhes desses dados serão apresentados nas subseções seguintes.

2.1. Dados de sensoriamento remoto

O campo de sensoriamento remoto tem crescido ao longo dos anos em variedade de sofisticação para o monitoramento da superfície da Terra (Rees, 2006). Às vezes, o sensoriamento remoto por meio de fotografias aéreas (Jordan et al., 2005) e / ou imagens de satélite é o único método disponível para estudar as geleiras, especialmente devido à inacessibilidade das áreas (Bolch e Kamp 2006). As imagens de satélite podem ser usadas para monitorar e medir a geleira, que é um bom substituto para estudar mudanças de balanço de massa em um longo período de tempo (Hall *et al*., 1987; Paul, 2000; Silverio e Jaquet 2005). A perícia para incluir /excluir um pixel na /da classe 'geleira' implicará na precisão do mapeamento da mesma (Williams *et al*., 1997; Vuille *et al*., 2008). Surazakov e Aizen (2006) usaram com sucesso uma combinação de imagens SRTM (*Shuttle Radar Topographic Mission*) e outros dados topográficos gerados a partir de modelos digitais de elevação (MDE) para a estimativa das variações volumétricas das geleiras. A perda de massa e elevação das geleiras está inversamente relacionada, quanto maior a elevação, menor será a perda de massa (Rabatel *et al*., 2013).

Imagens de satélite de várias fontes foram utilizadas nesta pesquisa. As imagens Landsat utilizadas estão disponíveis já ortorretificadas, portanto nenhuma correção geométrica foi necessária durante o pré-processamento das imagens. Estas imagens são baixadas em formato GeoTIF. As imagens Landsat e EO1 ALI podem ser baixadas do USGS, imagens IRS LISS III (para o período de 2012, por causa de descontinuidade de imagens de Landsat) estão disponíveis gratuitamente do INPE (Instituto Nacional de Pesquisas Espaciais). As imagens de ASTER podem ser obtidas a partir NASA Reverb depois de registrar um projeto.

Além de imagens multiespectrais, foram utilizados nesta pesquisa os modelos digitais de elevação (MDE) a partir de modelos digitais de elevação ASTER globais (GDEM) também para calcular ALN, pois possuem uma resolução espacial de 30 m. As morainas das geleiras são cobertas por neve durante a temporada de alta precipitação e a ablação será maior durante a época de baixa precipitação. É difícil calcular a área da geleira ou ALN a partir de imagens de satélite durante essas épocas de maior precipitação e, portanto, todas as imagens utilizadas neste estudo foram selecionadas por não possuírem cobertura de neve excessiva. Além disso, o co-registo das imagens multiespectrais e dos DEM foram realizados, bem como correções radiométrica das imagens antes da aplicação de outros algoritmos de processamento. Todas as imagens foram corrigidas com base no ângulo zenital solar. As etapas de processamento de imagens foram realizadas usando os pacotes de software Erdas Imagine e ESRI ArcGIS 10.1.

2.2. Dados meteorológicos

Dependendo do local de estudo, os dados meteorológicos utilizados são de fontes diferentes. A Universidade de Delaware disponibiliza para acesso o público, dados reticulados de precipitação e temperatura mensal (acima de 2 m do nível do solo) com uma resolução horizontal de 0,5º x 0,5º lat-long durante o período de 1948-2008. Estes dados foram obtidos a partir de um grande número de estações globais, incluindo a Rede Global do Clima Histórico (GHCN2) e os arquivos de Legates e Willmott (1990).

Dados de NCEP/NCAR Reanalysis1 (Kalnay *et al.*, 1996) tem uma resolução de 2,5º x 2,5º lat-long e os dados disponíveis dos anos de 1948-2014 foram também usados nesse projeto. Apesar da baixa resolução em comparação com outros dados, os dados do NCEP/NCAR são amplamente utilizados nas regiões de maior altitude nos Andes (Schauwecker *et al.*, 2014). Para calcular a anomalia de temperatura em alguns locais de estudo durante 1950-2012 foram utilizados os dados da Unidade de Pesquisa Climática (CRU) da Universidade de East Anglia (New *et al.*, 2002) com uma resolução de 0,5º x 0,5º lat-long. Esses dados têm boa resolução, mas a cobertura é esparsa se comparados com os dados do NCEP/NCAR ou Delaware. No entanto, estes dados têm uma boa cobertura na Cordilheira Branca e, portanto, foram utilizados neste estudo. Dados de ERA-Interim tem uma resolução de 0,5º x 0,5º lat-long e os dados disponíveis durante 1979-2014 também foram usadas nesse projeto, para entender as tendências recentes.

O Índice de Niño Oceânica (ONI) é um dos índices primários utilizados para monitorar o El Niño - Oscilação Sul (ENOS). A ONI é calculada pela média de anomalias de temperatura da superfície do mar em uma área do Oceano Pacífico equatorial leste-central, que é chamada região Niño-3.4 (5ºN a 5ºS; 170ºO a 120ºO). Além disso, a média de tempo de 3 meses é calculada de modo a isolar melhor a variabilidade intimamente relacionada com o fenômeno ENOS. O Centro de Previsão Climática da NOAA determinou a temperatura média mensal da superfície do mar para uma faixa específica do Oceano Pacífico tropical, pela média de medições recolhidas ao longo dos últimos 30 anos. Os Índices Oceânicos Niño (ION) podem ser baixados do site do Centro de Previsão Climática (CPC), *National Oceanic and Atmospheric Administration* (NOAA). Os valores da região Niño 3.4 foram utilizados neste estudo. Nesses dados, os episódios quentes e frios foram definidos quando um limite de +/- 0,4ºC é atingido por um período mínimo de seis temporadas consecutivas sobrepostas. Sabe-se que a temperatura média da superfície do mar aumentou no período de 1975-2010 (Trenberth, 1997).

A Oscilação Decenal do Pacífico (ODP) é um índice com base na variação do TSM no norte do Pacífico e pode ser baixado no site do *Joint Institute for the Study of the Atmosphere and Ocean* (JISAO) ou Agência Meteorológica do Japão. Os valores padrão do índice de ODP são derivados das anomalias mensais de TSM no Oceano Pacífico norte a partir de 20ºN. Ao contrário de ENOS, os eventos ODP podem persistir por várias décadas. Por exemplo, uma fase de aquecimento contínuo no período de 1925 a 1946 e uma fase fria de 1947 a 1976. De 1977 a 1998, outra fase de aquecimento de 21 anos ocorreu. No entanto, esses ciclos decenais foram recentemente divididos: no final de 1998 a ODP entrou em uma fase fria que durou apenas quatro anos seguidos de uma fase de aquecimento de 3 anos de 2002 a 2005, neutro até agosto de 2007 e de repente mudou para uma fase fria que durou até 2013, quase 6 anos com apenas uma curta interrupção durante o El Niño moderado no outono/inverno 2009-2010 (NOAA Fisheries, 2014). Desde o início de 2014, no entanto, a ODP mudou de fase novamente e tem sido fortemente quente (NOAA Fisheries, 2014).

3. METODOS

A metodologia é dividida em duas partes. Na primeira etapa as imagens de satélite foram utilizadas para calcular a ALN durante a estação seca, e por sua vez esta ALN pode ser utilizada para estimar a altitude da linha de equilíbrio do ano. Na segunda etapa, as cinco variáveis meteorológicas aqui utilizadas foram analisadas e depois correlacionadas com a linha de equilíbrio calculada e com as anomalias nas variáveis meteorológicas.

3.1. Aplicação de sensoriamento remoto para estimar a ALN nos Andes tropicais

Um grande número de trabalhos de pesquisa sobre a medida de características físicas e da superfície das geleiras está disponível, incluindo posição da linha de neve e altitude, elevação da superfície e posição de término usando sensoriamento remoto e fotogrametria (Aniya *et al*., 2000; Arnaud *et al*., 2001; Bamber e Rivera 2007; Rabatel *et al*., 2012; Veettil, 2012). A série Landsat (MSS, TM, ETM+ e OLI) possui um acervo histórico de imagens desde 1972, e estão disponíveis para acesso público. As imagens livres de nuvens tomadas durante o fim dos meses de verão foram utilizadas com a finalidade de diminuir a dificuldade na delimitação da margem de gelo devido à ablação excessiva. Os comprimentos de onda (bandas) do visível e do infravermelho são utilizados de forma eficaz no monitoramento da linha de neve em condições livres de nuvens (Arnaud *et al*., 2001).

A identificação de neve e gelo usando sensoriamento remoto na teoria é fácil, mas na prática não é tão simples (Albert, 2002). Isto porque a refletância espectral da neve e do gelo varia de acordo com a quantidade de impurezas e de água de degelo acima do gelo e se altera com o envelhecimento da neve/geleira. **Figura 10** mostra as curvas de refletância espectral do gelo e da neve. Dois sistemas de satélites utilizados nas últimas 2-3 décadas, Landsat e SPOT, sofrem limitações na resolução espacial e espectral, respectivamente. Uma das desvantagens do uso de imagens de satélite é que pode ser necessário usar diversos algoritmos dependendo do lugar ou da época, ou seja, algoritmos específicos de lugar para lugar e/ou de época para época. Métodos que utilizam imagens Landsat variam desde composições coloridas/falsa-cor (combinações de bandas e de razão de bandas), passando por análise de componentes principais e indo até índices para orientada a objetos na imagem. A fim de garantir a precisão e reduzir os erros, as imagens foram atmosfericamente corrigidas e co-registadas antes do processamento adicional.

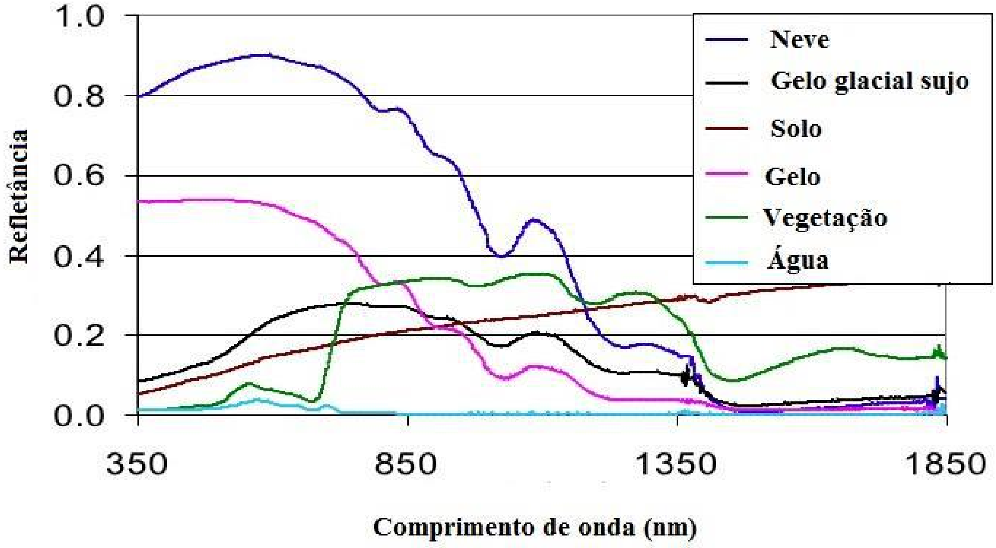

Figura 10—Curvas de refletância espectral do gelo e da neve (Fonte: Kulkarni et al., 2007)

Um dos métodos mais eficazes para mapear as geleiras é a delimitação manual, mas, neste caso, não se explica devido ao grande número de imagens, mais de 30 anos. Nesta pesquisa, foram utilizadas as variações de ALN para compreender as variações anuais e as mudanças na área da geleira por mudanças decenais. Deve-se assegurar que as imagens foram adquiridas no final da estação seca e que estão livres de nuvens. A área das geleiras pode ser calculada através da aplicação de um limiar para as imagens de diferença normalizada de índice de neve (NDSI). As imagens de NDSI são geralmente calculadas a partir do verde (TM2, 0,52 µm - 0,60 µm) e do infravermelho médio (TM5, 1,55 µm - 1,75 µm) das bandas Landsat utilizando a seguinte equação (**Eq. 1**): -

$$NDSI = [(TM2-TM5) / (TM2+TM5)] \qquad \textbf{Eq. 1}$$

O valor limite a ser aplicado para delinear a margem da geleira pode variar de lugar para lugar e até de imagem para imagem (Wang e Li, 2003). Os métodos de NDSI têm sido amplamente utilizados por muitos pesquisadores na Cordilheira Branca (Silverio e Jaquet, 2005; Racoviteanu *et al*., 2008).

As mudanças na área podem nem sempre ser um bom indicador das variações anuais no clima devido às variações interanuais na cobertura de neve. No entanto, ALN é um bom indicador de variações climáticas anuais, e estas podem ser calculadas com base em Rabatel *et al*. (2012). A maior ALN calculada no final da estação seca (maio-agosto) pode ser tomada como um representante da linha de equilíbrio (LE) (Rabatel *et al*., 2012), particularmente nos trópicos externos e subtrópicos. Com base na comparação e validação com os dados de campo no Glaciar Zongo na Bolívia e Glaciar Artesonraju no Peru, verifica-se que a ALN é um bom representante para LE e, portanto, pode ser usada para medir as mudanças de equilíbrio de massa anuais (Rabatel *et al*., 2012). Em contraste com os trópicos internos, há uma forte sazonalidade de precipitação nos trópicos externos e devido a esta característica, podemos usar o maior valor de ALN durante a estação seca como representante de LE do ano. Para calcular ALN a partir de imagens do satélite Landsat, a composição falsa-cor R5-G4-B2 foi usada com certo limiar aplicado a TM2 e TM4. A fim de obter boa precisão para a ALN, o limiar aplicado a TM4 pode variar entre 60 e 135 e para TM2 varia entre 80 e 160. Deve-se notar que as imagens com uma resolução radiométrica de 16-bit foram convertidas em 8-bit para aplicar o mesmo algoritmo. Exemplos de vários métodos para calcular a altitude da linha de neve são mostrados na figura 11.

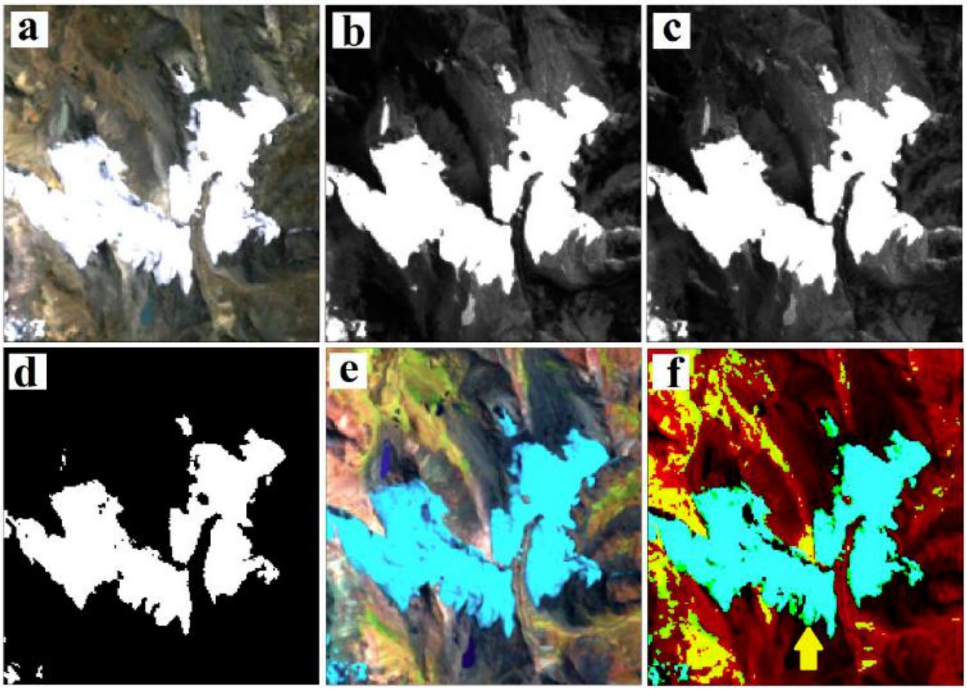

Figura 11—Teste de combinações diferentes de bandas e razão de bandas aplicadas em imagem de Landsat-5 adquiridas em 14 de agosto de 1991 facilita a identificação da linha de neve: (a) combinação de bandas espectrais 3, 2 e 1; (b) relação de 3/5; (c) relação de 4/5; (d) a diferença normalizada de índice de neve (NDSI), com limite de 0,6; (e) combinação de

bandas espectrais 5, 4 e 2; (f) igual a (e) com o limiar de 120 e 135 para as bandas 4 e 2, respectivamente. Em (f) a seta amarela indica a posição da linha de neve no Nevado Tuco.

3.2. Cálculo das anomalias na precipitação e temperatura

As anomalias na precipitação e na temperatura foram calculadas subtraindo-se o valor anterior do valor atual (valores mensais) e, em seguida, representadas graficamente através do MATLAB. Essas anomalias foram correlacionadas com as mudanças na ALN anual e a freqüência de ocorrência de ENOS e ODP. Para os dados da Universidade de Delaware e NCEP/NCAR Reanalysis1, apenas uma célula da grade é usada para calcular as anomalias. A média de quatro células da grade foi utilizada para calcular as anomalias usando dados ERA-Interim e CRU. Para evitar discrepâncias entre vários dados, os dados do NCEP/NCAR Reanalysis1 foram aplicados em todos os lugares de estudo para conferir se os outros dados também apresentavam tendências similares.

3.3. Estimando a teleconexão entre as variações de ALN e ENSO/ODP nos locais de estudo

A análise de séries temporais é uma das melhores maneiras de entender a teleconexão entre as ocorrências de ENSO (ODP) e as variações de linha de neve. Deve-se notar que há uma defasagem de 1 a 3 meses entre a ocorrência de El Niño/La Niña e as respectivas variações nos parâmetros meteorológicos do continente, dependendo da distância geográfica da costa do Pacífico (Kumar and Hoerling, 2003). Cada estação chuvosa inicia a linha de neve, mas durante a estação seca, a ALN não é estável em relação à sua posição no final da estação chuvosa e isso precisa ser levado em conta. Mesmo que a ablação, devido ao derretimento, seja inexistente em decorrência das temperaturas frias na estação seca, a sublimação ainda está presente e é ainda maior do que na época das chuvas, devido às condições de céu predominantemente claros e a presença de uma gradiente vertical de umidade elevado. As mudanças na ALN de cada ano devem ser analisadas com a ocorrência de fases quentes ou frias de ENOS e com os regimes quentes e frios do ODP.

3.4. Erros no cálculo ALN utilizando imagens multiespectrais

Várias fontes de erro podem existir no cálculo da ALN. Devido às temperaturas muito baixas entre os diferentes locais de estudo, a presença de gelo limpo é rara, uma vez que é freqüentemente coberta de neve. Isso impede a detecção da área de ablação na parte inferior, o que é bastante incomum para uma língua clássica de geleira. A ALN calculada também depende muito da resolução do MDE usado e o co-registro de todas as imagens e do MDE é necessário a fim de comparar as ALN. Devido à topografia extremamente rugosa e ao aspecto variável em algumas partes dos locais de estudo, ambos influentes no cálculo da ALN, decidiu-se, portanto trabalhar em regiões específicas das geleiras, em vez de considerar a ALN em toda a região. As zonas de estudo selecionadas estão em encostas suaves onde a maior iluminação solar ocorre no inverno. O erro na estimativa da ALN é muito difícil de ser determinado, uma vez que depende do erro de co-registro da imagem em relação à resolução horizontal e vertical da MDE, mas também depende da inclinação do terreno no local. Em um terreno plano, um erro horizontal não teria qualquer impacto na determinação de altitude, já em um terreno com uma inclinação de 45°, um erro horizontal implicaria na mesmo magnitude de erro na altitude. Os erros verticais resultantes, no entanto, ainda estão de acordo com a precisão necessária para este trabalho, isto porque todas as geleiras consideradas nesta pesquisa possuem declividades menores (< 20°) e a precisão vertical de ASTER GDEM (V2) está dentro do escopo deste valor.

Uma das dificuldades na estimativa da linha de equilíbrio com base em ALN nos trópicos internos (Equador) é o fato de que existem duas estações de alta precipitação por ano e, portanto, não se mostra tão eficaz. Esta diferença entre a precipitação média mensal nos trópicos internos e trópicos externos está indicada na **figura 3** na revisão da literatura. Portanto, é utilizada a linha de neve máxima calculada durante a época baixa precipitação.

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CAPITULO IV

Combined influence of PDO and ENSO on northern Andean glaciers: a case study on the Cotopaxi ice-covered volcano, Ecuador

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Climate Dynamics 43, 3439-3448, 2014

Clim Dyn (2014) 43:3439–3448
DOI 10.1007/s00382-014-2114-8

Combined influence of PDO and ENSO on northern Andean glaciers: a case study on the Cotopaxi ice-covered volcano, Ecuador

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Received: 11 November 2013 / Accepted: 6 March 2014 / Published online: 22 March 2014

Abstract This paper describes the application of remote sensing in monitoring the fluctuations in one of the mountain glaciers in the Ecuadorean Andes during the past few decades using ASTER, EO-1 ALI, Landsat MSS, TM and ETM + images. Satellite images were used to calculate the snow line altitudes (SLAs) during the period 1979–2013. Cotopaxi ice covered volcano was studied as representative of Ecuadorian glaciers in the eastern cordillera. Precipitation and air temperature data from various gauging stations within the range of 30 km from the study site and monthly discharge and water level data from a gauging station were also utilized in this study. Anomalies in precipitation and temperature were found to be slightly different in the Cotopaxi region compared to nearby Antizana in the same cordillera and Chimborazo region in the western cordillera. An attempt to correlate the El Niño—southern oscillation phenomenon with the glacier fluctuations in Ecuadorian Andes was done successfully. Cold and warm regimes of Pacific Decadal Oscillation is also considered. The calculated glacier fluctuations obtained were similar to that performed on the nearby Antizana 15 in the eastern cordillera during 1995–2002. Precipitation and temperature anomalies were similar with Antizana 15. It is evident from the research that SLAs were highly fluctuated between the period of occurrence of El Niño and La Niña events. It is also seen that the glacier fluctuations show a negative mass balance trend in during the warm regime of Pacific Decadal Oscillation during the past three decades. Glaciated areas were advanced during the La Nina events in the cold regime of PDO during 1998–2002.

Keywords Andean glaciers · ENSO · PDO · Mountain glaciers · Cotopaxi · Precipitation

1 Introduction

Glaciers, particularly mountain glaciers and ice caps in the tropical region, are extremely sensitive to environmental fluctuations and considered as useful indicators of climate change (Barry 2006). The response of glaciers to environmental changes is complex (Oerlemans 1989). Retreat of mountain glaciers and ice caps were found to be linked directly with historical temperature records (IPCC 2001). Majority of the population living in the surrounding area of tropical mountain glaciers (where seasonal precipitation is almost absent) depends on the glacier melt water for their everyday fresh water and hydroelectric energy needs. If these glaciers are going to be disappeared forever, it will have serious consequences on the water availability to such a massive population. In recent decades, our knowledge about glacier fluctuations has been improved significantly due to the understanding of modern climate-glacier relationships as well as due to rapid development of

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technologies and concepts in Paleoclimatology (Solomina et al. 2008). Global warming scenarios have tend to invoke images of rising sea level, increased melting of the World's glacier bodies and accelerated calving rates at ice-sea margins (Benn and Evans 1998), even though some debris-covered Himalayan glaciers were found to be showing an exceptional positive mass balance and advances (Veettil 2009, 2012). In other words, glaciers are good proxies for monitoring climate change (Kaser et al. 2003b).

Andean glaciers in Ecuador, Peru, Bolivia and Chile were found to have undergone an interrupted retreat after the ''Little Ice Age'' (Kaser 1999; Kaser et al. 2003a; Rivera et al. 2005; Vuille et al. 2008a), especially those laying in lower altitudes (Kaser 1999). Knowledge about the complex behavior of tropical glaciers is rare compared to that in the polar region (Kaser 1999). About 99.7 % of tropical glaciers are situated in South America and the rest are distributed in Africa and Asia (Kaser 1999). Many researchers have tried to find a connection between this rapid and discontinuous glacier ablation and the occurrence of El Niño and La Niña events and/or anthropogenic climate change (Beniston et al. 1997; Favier et al. 2004a, b; Francou et al. 2004; Haeberli et al. 2000, 2007; Jomelli et al. 2009; Rabatel et al. 2005; Vuille et al. 2008a, 2008b). In mountain environments, such as the Ecuadorian Andes, prevailing ablation energy in glacier retreat is provided by longwave radiation, which penetrates through cloudy skies (Sicart et al. 2010). Melt rates of tropical glaciers are sensitive to near surface vapor pressure, atmospheric humidity effects, sublimation, annual precipitation rates (Vuille et al. 2008a), air temperature and albedo (Kaser et al. 2003a). On an interannual timescale, precipitation variability is linked with atmospheric circulation and is the driving mechanism in glacier mass balance fluctuations (Vuille et al. 2008a). Some ice-covered volcanoes such as Chimborazo and Carihuairazo are having a complicated volcanic history and were identified with four generations of glacial deposits (Smith et al. 2005). Ice core records (Ginot et al. 2010) and radiocarbon dating (Rodbell 1992; Hajdas et al. 2003; Jomelli et al. 2009) helped the scientific community to understand the complex history of tropical glaciers even though it is difficult to obtain organic materials trapped inside glacial landforms (Benn et al. 2005). Other methods include lichenometry (Rodbell 1992; Rabatel et al. 2005; Jomelli et al. 2009), remote sensing and photogrammetry (Haeberli et al. 2003) to calculate glacier fluctuations.

El Niño—southern oscillation (ENSO) is a global scale ocean–atmosphere phenomenon which causes climate variability on interannual time scales, with irregular fluctuations between its warm phase (El Niño) and cold phase (La Niña) with a periodicity of 2–7 years (Garreaud et al. 2009). ENSO is normally associated with sea surface temperature (SST) changes (Joseph and Nigam 2006) and hence can be monitored based on SST variations. Precipitation and atmospheric circulation anomalies in South America (Grimm et al. 2000) and Indian monsoon rainfall (Ropelewski and Halpert 1987) were found to have strong relationship with the occurrence of ENSO events. Aceituno (1988) studied the patterns of correlation between Southern Oscillation in South America and surface pressure, wind, temperature, precipitation and other hydro-meteorological indices and found that Southern Oscillation controls mainly the summer time climate variability in South American tropics. Opposite rainfall and temperature anomalies were observed during the El Niño and La Niña events (Garreaud et al. 2009) which show its impact on the tropical climate. El Niño is associated with higher temperature, lower precipitation, wind speed and albedo while La Niña events are with higher precipitation rates, wind speed and albedo and lower temperature in the tropical Andes (Chevallier et al. 2011). There are some controversial hypothesis connecting the occurrence of ENSO with volcanic forcing (Adams et al. 2003), which in turn affects the ice caps on the active volcanoes, though these are temporary and local in occurrence. The Pacific Decadal Oscillation (PDO) is a climate index based on the North Pacific SST variations (Mantua et al. 1997) with warm (positive index) and cold regimes (negative index). Warm and cold regimes can persist for several decades. Since 1998, these decadal cycles have broken down. PDO can modulate the interannual relationship between the ENSO and global climate and can cause interdecadal climate variability in the tropical Pacific (Mantua et al. 1997). The focus of this research is on the combined influence of ENSO and PDO on glacier variations in the Ecuadorian Andes.

2 Study site

Cotopaxi, located about 60 km southeast of the capital Quito, is one of the highest volcanoes in the world and is the second highest summit in Ecuador (5,897 m a.s.l, 0°40′S, 78°25′W) (Fig. 1). This extensively ice-covered volcano in Ecuador and its fluctuations in mass balance during the last 30 years is studied here. Glaciers in Ecuador are situated in the inner tropics and are located in two cordilleras—Cordillera Occidental or Western Cordillera (Lat 0°22′N–1°29′S; Lon 78°20′W–78°48′W) and Cordillera oriental of Eastern Cordillera (Lat 0°1′N–2°20′S; Lon 77°54′W–78°33′W). Mt. Cotopaxi is situated in the Cordillera Oriental and is the highest active volcano in the world. Eastern cordillera is directly exposed to moist easterly winds from the Amazon basin. There are two precipitation maximums per year–March, April and May (main) and September, October and November

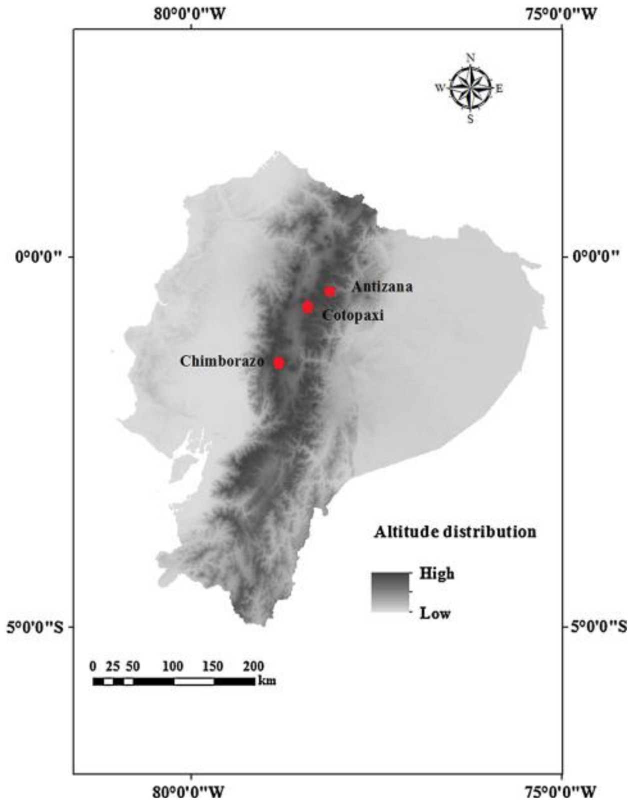


Fig. 1 Location of Cotopaxi and nearby Antizana and Chimborazo volcanoes and altitude distribution in the Ecuador

(secondary). Seasonal variations in in temperatures are not significant but the interannual variability is considerably large and wind is the principal factor of seasonality in the eastern cordillera in Ecuador (Francou et al. 2004). In Ecuador, precipitation is modified by the Andean mountain system, mesoscale phenomenon of the valley winds and the ocean currents (ENSO and Humboldt current) (Bendix and Lauer 1992) which in turn affects the mass balance changes. Air temperature is another factor which controls mass-energy balance in this region and hence is more sensitive to climate variations than in the outer tropics (Favier et al. 2004a, b; Jomelli et al. 2009). Incoming shortwave radiation is maximum during the periods close to equinox (March–April and September) and the absence of solid precipitation during this period lead to significant melt rate (Rabatel et al. 2013). Recent eruptions reported were in 1742–1744, 1768 and 1877 (Jordan et al. 2005). (Cáceres et al. 2004) tried to quantify the ice discharge that may occur in the future due to laharitic phenomenon caused by volcanic eruptions. About 4 % of the World's tropical glaciers are situated in the Ecuadorian Andes (Kaser 1999). Andes is the runoff dividing line between the Pacific Ocean and the Amazon basin with a north–south orientation (Laraque et al. 2007). Nearby Glaciated volcano Antizana 15 (in the eastern cordillera, between 5,760 and 4,840 m a.s.l) above the Amazon basin had undergone a massive retreat from 1956 to 1998 (Francou et al. 2000). Antizana 15 were reported to be advanced between 1999 and 2001 due to extensive La Niña conditions prevailed in the region (Francou et al. 2000, 2004). Chimborazo (6,267 m a.s.l) was reported to be lost 21 % of its glaciated area between 1986 and 2012 (Frenierre 2012) but in this study, a multitemporal approach to calculate annual changes was not done. This research is trying to find a correlation between the occurrence of ENSO (and PDO) and the mass balance changes on the ice-covered Cotopaxi volcano based on remote sensing and meteorological datasets. Meteorological conditions in the nearby Antizana and Chimborazo regions were also considered in this study. Figure 2 shows the three outlet glaciers towards the north of Cotopaxi used in this study.

3 Datasets

This paper is the outcome of an interdisciplinary research based on multisource data derived from remote sensing, meteorology and hydrology. Following subsections describe the datasets used and their suitability in the research.

3.1 Remote sensing data

The field of remote sensing has been growing over years in a range of sophistication for monitoring the Earth surface (Rees 2006). Sometimes aerial photographs (Jordan et al. 2005) and/or remote sensing is the only method available to study glaciers, especially due to inaccessibility in remote areas (Bolch and Kamp 2006). Satellite images can be used for monitoring glacier extent, which is a good proxy to study mass balance changes for a long period of time (Hall et al. 1987; Paul 2000; Silverio and Jaquet 2005). The expertise to include/exclude a pixel into/from the class 'glacier' would decide the accuracy of glacier mapping (Williams et al. 1997; Vuille et al. 2008a). Surazakov and Aizen (2006) used a combination of SRTM images and other topographic data generated from digital elevation models (DEMs) for the estimation of volumetric changes of mountain glaciers successfully. Mass loss and elevation of glaciers are directly related—higher the elevation, lower will be the mass loss (Rabatel et al. 2013). Selection of images with suitable spatial and spectral resolution and type of the glacier to be monitored are crucial factors for delineating glacier margins efficiently (Paul 2000). It is now possible to calculate ice thickness distribution of mountain glaciers using the GLIMS database (Huss et al. 2011). GLIMS provides two-dimensional information about mountain glaciers and DEMs on a global scale which are comparable to field data.

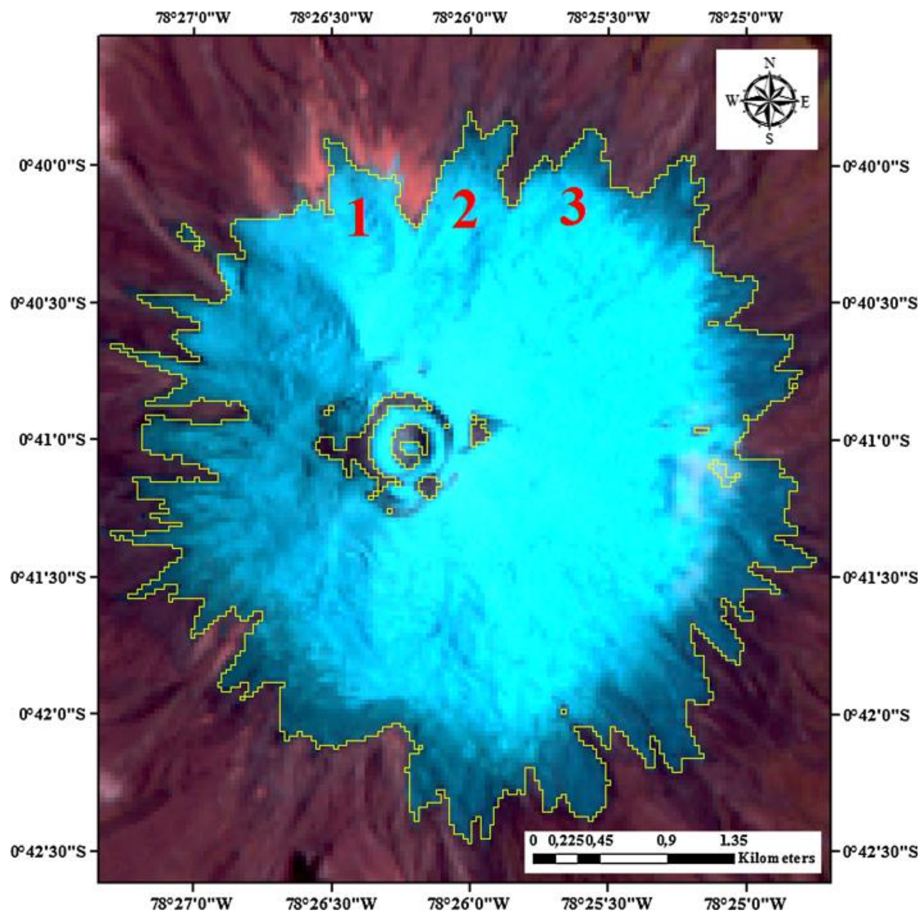


Fig. 2 Outlet glaciers selected for calculating SLA

Due to the abundance of multispectral images from Landsat series (MSS, TM, ETM+ and LDCM), EO-1 ALI and Terra ASTER, it is now possible to calculate multitemporal glacial changes. These images from 1970s to present are now available free of cost. Landsat ETM+ and EO1 ALI sensors are providing panchromatic images as well with a spatial resolution of 15 and 10 m respectively. Glacier moraines would be covered by snow during winter and ablation rates would be higher during the summer. It is difficult to calculate the exact glaciated area from satellite images during high precipitation season and hence a few images in the summer devoid of cloud cover were used here. There are two precipitation seasons per year in the study site (March–May and September–November). Digital elevation model (DEM) derived from ASTER global digital elevation models (GDEM) was also used for calculating the snow line altitude (SLA). All the image processing steps were done using IDRISI Selva and ESRI ArcGIS 10.1 software packages. Meteorological data were processed and plotted used MATLAB.

3.2 Meteorological, hydrological and climatological data

Monthly mean precipitation (MMP) varies in Ecuador. In the Cotopaxi region MMP is different from that in the Chimborazo (Cordillera Occidental) and slightly different from the Antizana. Figure 3 demonstrates how the MMP (1976–2008) varies in Cotopaxi (on the Altiplano), Antizana (just above the Amazon basin) and Chimborazo (more to the west, above the Pacific shoreline). Monthly global gridded high-resolution precipitation and temperature (above 2 m from the ground level) station data with a resolution of 0.5° lat-long during a period of 1979–2008 from the University of Delaware is used in this research. These data were derived from a large number of stations including Global Historical Climate Network (GHCN2) and archive of Legates and Willmott.

Monthly discharge data and monthly average water level data from a hydrological gauging station (Station: Toachi AJ Pilaton, Code: H161, Elevation: 820 m, Latitude: 0°18′51′′ N, Longitude: 78°57′12′′W) which is situated within the 30 km of the Cotopaxi Volcano were also used for this research. This data is freely available from the National Institute of Meteorology and Hydrology (INAMHI), Ecuador.

Ocean Niño indices (ONI) and PDO index were downloaded from National Oceanic and Atmospheric Administration (NOAA 2013) climatic prediction center. In this data, cold and warm episodes were defined when a threshold of ±0.5 °C in the Niño 3 region (0.4 °C in the

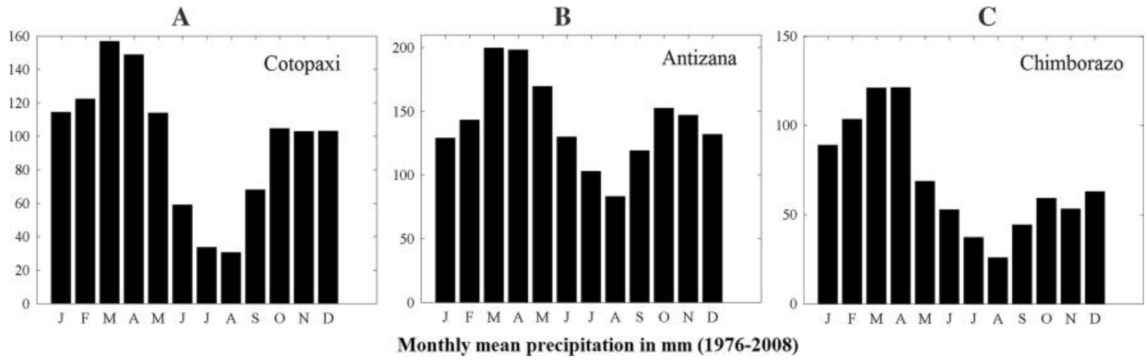


Fig. 3 Different monthly mean precipitation in millimeters between observed 1976 and 2008 in **a** Cotopaxi, **b** Antizana and **c** Chimborazo

Niño 3.4 region) is met for a minimum of five consecutive months (Trenberth 1997). It is seen that the average Pacific SST has been increased during 1975–2010.

4 Methodology and results

The entire research was done in three sections. Firstly, multitemporal glacier mapping was done using satellite images. SLA was calculated for selected outlet glaciers of Cotopaxi, based on the availability of cloud free images. Secondly, anomalies in precipitation and temperature were calculated from 1979 to 2008 using the gridded datasets. Changes in the river discharge volume and water level were also analyzed during this step. Finally an attempt to correlate the changes in the SLAs with the above anomalies and ONI and PDO indices.

4.1 Glacier changes during 1979–2013

Because of the interannual variability of snow cover extent, SLA can be used as a more accurate climate indicator rather than glacier area (Arnaud et al. 2001). In Ecuador, mean ablation rate is almost constant throughout the year on seasonal time scales (Favier et al. 2004b). The interannual mass balance variability is controlled by annual variations in air temperature (Francou et al. 2004). A study based on SLA on Sajama ice-covered volcano in Bolivia (Veettil et al. 2013) could successfully correlate the glacier changes with ENSO though a surface area based test did not give acceptable results. PDO index was not considered while interpreting the influence of ENSO in Veettil et al. (2013). Highest SLA detected using satellite imagery towards the end of summer can be taken as a proxy of the equilibrium line altitude (ELA) of the year (Rabatel et al. 2012). Various glacier mapping algorithms were tried to get a proper outlet SLA because images of different spatial and spectral resolutions were used. Manual delineation was applied when (1) Landsat MSS images were used

Table 1 SLA from 1979 to 2013

Date of Acquisition	Sensor	SLA in m		
		Outlet 1	Outlet 2	Outlet 3
04.02.1979	Landsat MSS	4,801	4,684	4,654
13.11.1987	Landsat TM	4,858	4,744	4,717
08.06.1988	Landsat TM	4,860	4,738	4,717
10.11.1989	Landsat TM	4,878	4,772	4,732
15.10.1991	Landsat TM	4,904	4,769	4,736
12.10.1996	Landsat TM	4,805	4,749	4,727
18.12.1997	Landsat TM	4,918	4,786	4,741
29.10.1999	Landsat ETM+	4,941	4,787	4,741
07.10.2000	Landsat ETM+	4,903	4,777	4,741
03.11.2001	EO1 ALI	4,915	4,777	4,741
27.11.2004	Landsat ETM+	4,894	4,786	4,741
09.07.2005	Landsat ETM+	4,970	4,811	4,751
29.08.2006	ASTER	4,970	4,849	4,747
31.07.2007	Landsat ETM+	4,981	4,878	4,764
17.07.2008	Landsat ETM+	4,953	4,874	4,764
06.09.2009	Landsat ETM+	4,970	4,867	4,757
12.01.2010	Landsat ETM+	4,960	4,851	4,751
02.01.2012	Landsat ETM+	4,901	4,867	4,764
21.06.2013	Landsat 8	4,915	4,867	4,769

(2) images have weak cloud cover (3) the glacier was undergoing rapid ablation because some algorithms based on albedo characteristics were failed to get good results. Manual approach is dependent on the expertise of the researcher and is time consuming when a large number of multitemporal data to be processed (Paul 2000; Paul et al.

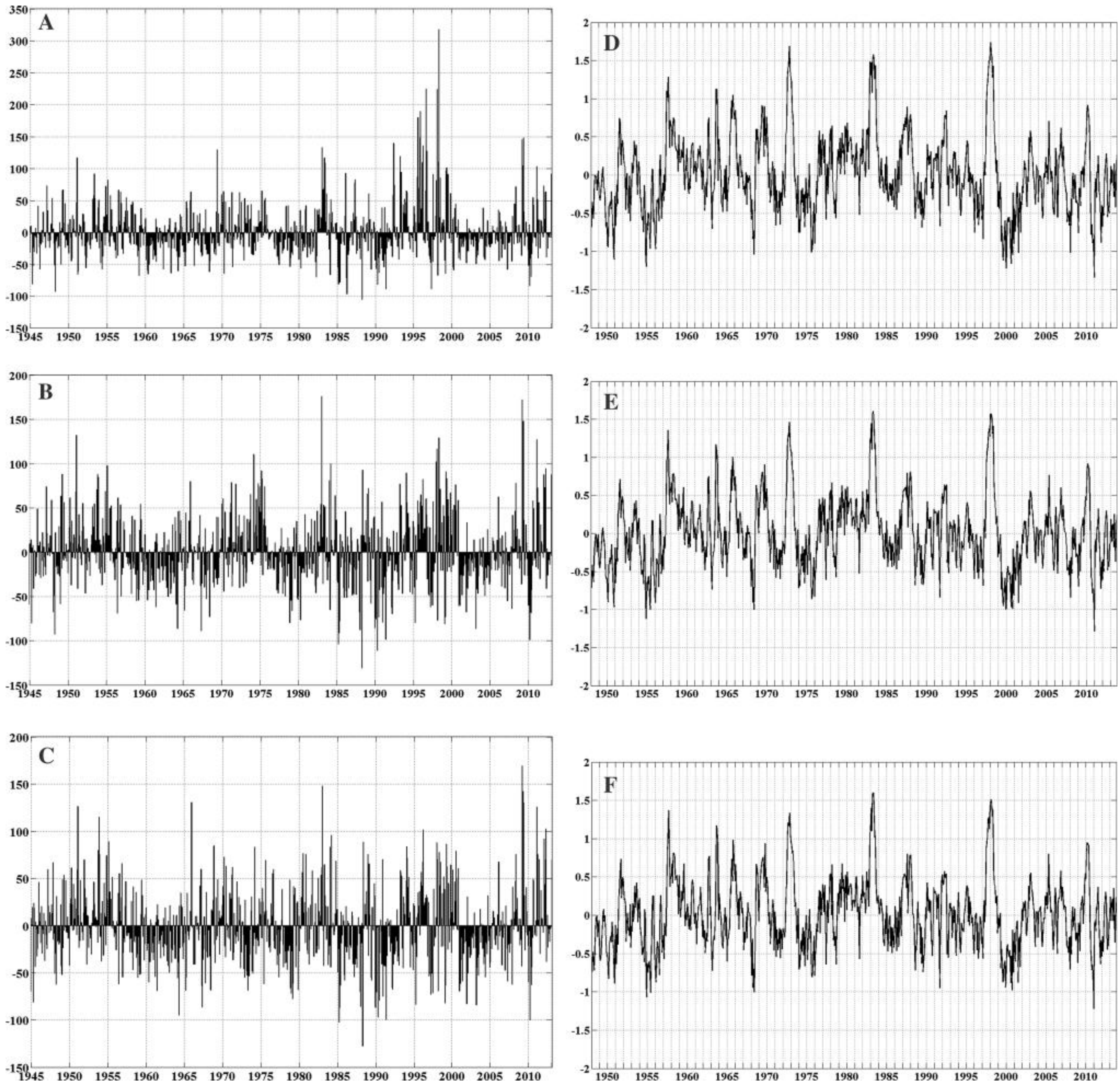


Fig. 4 Anomalies in precipitation (**a** Chimborazo; **b** Cotopaxi; **c** Antizana) and temperature (**d** Chimborazo; **e** Cotopaxi; **f** Antizana)

2004). Automated algorithms, which are used directly irrespective of the user expertise, were mainly used here. Glacier mapping by calculating normalized difference snow index (NDSI) and/or band ratios between visible and infrared channels and then applying a suitable threshold (Silverio and Jaquet 2005) were the automated methods tried here. NDSI method is suitable using Landsat TM and ETM + images. The choice of threshold can vary from place to place or images of same location with different dates of acquisition. Other methods include principal component analysis (PCA) and post-classification image analysis (Sidjak and Wheate 1999) which were found to be suitable in the case of less sampling areas. A combination of automatic and manual approach was found to be a suitable where rapid fluctuation of glaciated area occurs. SLA calculated from above methods are given in Table 1.

4.2 Anomalies in precipitation and temperature

Precipitation and temperature anomalies were calculated from the gridded datasets mentioned above in the Cotopaxi region using MATLAB. For a comparison, anomalies in

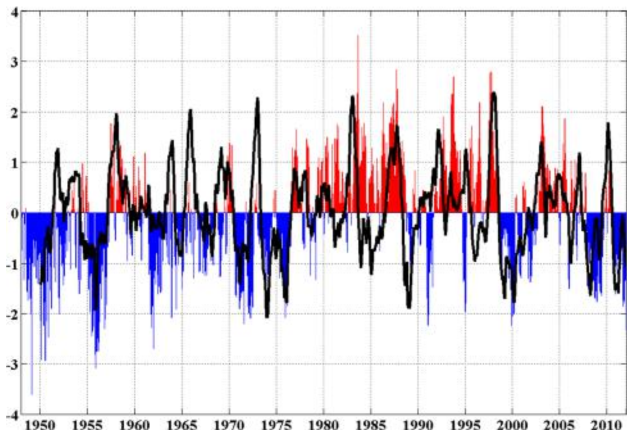
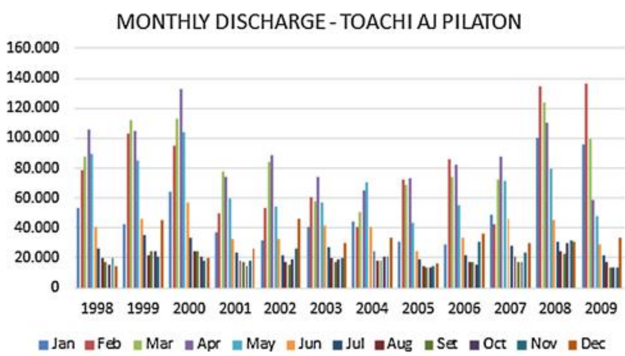


Fig. 5 Ocean Nino Index between 1948 and 2012 (*solid black*) and PDO index between 1945 and 2012 (*red* warm regime, *blue* cold regime)



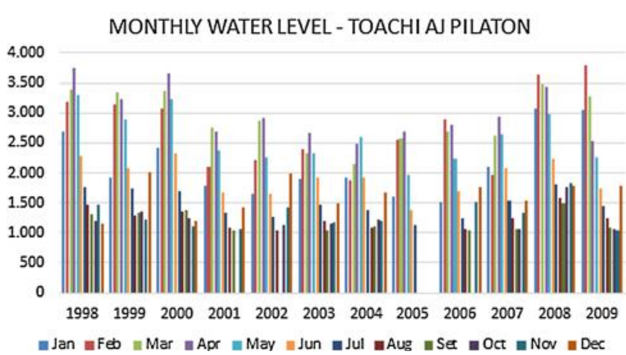


Fig. 6 Monthly discharge volume and water level from the nearest available station

precipitation and temperature in the Chimborazo and Antizana were also calculated (separately rather than finding a weighted average value in the Andean region). Figure 4 shows the anomalies obtained. Ocean Nino Index (ONI) and PDO time series data from NOAA is given in Fig. 5. It is seen that El Niño years were associated with higher temperature and lower precipitation. The precipitation anomaly in the case of Cotopaxi shows similar patterns of that in Antizana and is different from Chimborazo. ENSO events tend to peak around the end of the calendar year and hence the strongest El Nino events in 1982, 1991 and 1997 can be extended to 1983, 1992 and 1998 respectively. Monthly discharge volume and water level in the Toachi Aj Pilaton station is graphically summarized in Fig. 6.

We analyzed the MMP trends within the warmer and cooler regimes of PDO in the Cotopaxi, Chimborazo and Antizana regions separately. Warm regimes of PDO coincide with anomalously dry periods in the central and northern South Americas and anomalously warm temperatures in northwestern North America and northern South America (Mantua and Hare 2002). Different monthly precipitation rates were observed in these three regions. Antizana region, which is just above the Amazon basin, was found to be having higher precipitation compared to other two regions. Cotopaxi region on the Altiplano is having a higher rate of monthly precipitation than the Chimborazo (which is more influenced by the Pacific). Warm and cold regimes of PDO showed significant difference in precipitation.

5 Discussion and conclusion

Due to the interactions between global and local climatic factors, the anomalous behavior of tropical Andean glaciers is difficult to explain and not even good models exist to predict future changes, such as in the case of Greenland or Antarctica. Climate is the primary factor governing the size of the glacier and the mass balance varies with annual as well as long-term variations (Bennett and Glasser 2009). Previous studies suggested that the Cotopaxi glacier remained stable with no or little loss between 1956 and 1976 (Jordan et al. 2005). Jordan et al. (2005) also calculated that the Cotopaxi had lost 30 % of its glaciated surface area between 1976 and 1997, based on the study on selected outlet glaciers. Even though the SLAs were fluctuated with the cold and warm phases, it would not be sufficient to consider only ENSO to describe the climate variability on the glacier changes. It is mentioned in this study that the equatorial Pacific was abnormally warm during this period (between 1976 and 1997). From Fig. 5, it is clear that this warm phase is situated in the positive PDO during 1975–1997. A cooling trend is found during 1948–1975, 1998–2002 and 2008-present. The occurrence of El Niño became more frequent during the recent decades and the cold and warm regimes of the PDO has broken down. Nearby Antizana volcano, which is just above the Amazon Basin, lost its 33 % of its glaciated area between 1979 and 2007 (Rabatel et al. 2013). The Chimborazo volcano also lost its 57 % (from 27.7 to 11.8 km^2 of the glaciated area during 1962–1997 (Cáceres 2010). The average SST have been increased dramatically after 1975 and this would explain why the glacier has been retreated during 1976–1997. We suggest that if La Niña events occur during the warm regime of PDO, gain in the mass balance would be less compared to that occur during the cold regime of PDO. This would explain why the colder La Niña events could not contribute a positive mass balance during the recent PDO warm regime (1975–1997). The positive mass balance observed in the Antizana 15 by Francou et al. (2004) during 1999–2000 La Niña phase can be explained by the presence of a short PDO cold regime prevailed during 1998–2002 (1997–1998 was a strong El Niño period).

From this research, it is seen that even though there was an overall loss of the glaciated surface area, the retreat was not continuous but with a recent retreating trend (due to the warm PDO regime). This uneven change is tried to be explained in consideration with the ENSO. Francou et al. (2003) described how the mass balance is linked with ENSO phases along with seasonal cycles, where majority of mass balance variations occur during summer months. However, warm/cool anomaly of ENSO oscillation is found to affect glacier ablation after 4 months of its occurrence (Cadier et al. 2007). During 1984–1986 and between 1999 and 2001 La Nina phase was prevailing and the glacier has shown a positive mass balance. Long warm phases after 1986 and 2002 (El Nino + warm PDO) have been depleted the glaciated area that was gained during the previous La Niña period. El Niño years were associated with below-average precipitation in most of the cases (Vuille et al. 2000a). Exceptions were found which are probably due to local climate factors. In some cases in Ecuador and northern Peru, precipitation can be above normal during El Niño and below normal during La Niña (Mason and Goddard 2001). Precipitation rate and surface air temperature (and hence ENSO as well) are the mass balance controlling factors. The results obtained were similar with Wagnon et al. (2001) based on the study on Glacier Zongo, Bolivia during 1997–1998 El Niño year. Monthly mass balance and precipitation were found to be correlated during La Niña periods than El Niño periods (Francou et al. 2004). In addition, an increased dependence on ENSO phenomenon with mass balance mechanism is noticed in the recent years (2000s) compared to an earlier dates (1980s and 1990s) which shows that the glaciers are continuing disappearing in the present conditions possibly due to broken down PDO.

From the results obtained from this study, it is seen that the glacier fluctuations were dominated by ENSO phases and the glacier is disappearing gradually in correlation with the increase in recent SST variations. This recent increase in the Pacific SST is due to the warm regime of PDO and increased frequency of occurrence of El Nino. During El Niño years, the glacier had been undergone strong negative mass balance (Wagnon et al. 2001). This makes the tropical glaciers in a non-equilibrium condition under the present climate (Favier et al. 2004b). Even though there are limitations to use ENSO phenomenon for predicting future impacts, it helps to estimate climate anomalies (Mason and Goddard 2001) such as using contingency tables. Apart from small differences, all the ice caps studied showed similar response with the occurrence of ENSO. Another notable characteristic is that different ice caps showed different rates of response especially in different cordilleras. This means different calibration is needed in each case (Jomelli et al. 2009). There is a lag by 3 months in the zonal mean tropical anomalies relative to equatorial east pacific SSTs (Kumar and Hoerling 2003; Vuille et al. 2000b). In this study, a lag of 3 months was considered based on Vuille et al. (2000b) which clearly explains the lag in the changes in precipitation patterns and temperature. Temperature variability in the Ecuadorian Andes is dependent on the central equatorial pacific SSTs (Niño-3 and Niño-3.4). Small differences in precipitation anomalies found in the case of Chimborazo (western cordillera), Antizana and Cotopaxi (eastern cordillera) could explain how the ENSO effect varies spatially in the two different cordilleras. However, it is suggested that a research on glaciers in two different cordilleras from the outer and inner tropics would explain whether the ENSO and PDO influences are different in the two cordilleras or not. Significant difference in the behavior of glaciers to climate forcing and the seasonality of precipitation (and hence melting) were found in the inner and outer tropics based on a study on Zongo Glacier in Bolivia (Favier et al. 2004b). Wind speed or sublimation data were not used in this study. On a long-term basis, when combined with ice core records and lichenometric analysis, the picture will be more unambiguous.

Acknowledgments We would like to acknowledge FAPERGS (Fundação de Amparo à Pesquisa do Estado do Rio Grande do Sul), Brazil, for providing financial support and INAMHI (Instituto Nacional de Meteorologia e Hidrologia), Ecuador, for providing meteorologicl and hydrological data. We would like to thank Dr. Jefferson Cardia Simões, Centro Polar e Climático (CPC), UFRGS for providing infrastructure and software facilities. The first author acknowledges Ms. Geana Veiga Aurelio for her help with statistical data analysis.

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CAPITULO V

Recent trends in annual snowline variations in the northern wet outer tropics: case studies from southern Cordillera Blanca, Peru

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Theoretical and Applied Climatology, 2016

Theor Appl Climatol
DOI 10.1007/s00704-016-1775-0

ORIGINAL PAPER

Recent trends in annual snowline variations in the northern wet outer tropics: case studies from southern Cordillera Blanca, Peru

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Received: 18 November 2015 / Accepted: 13 March 2016

Abstract This paper describes the changes in the annual maximum snowlines of a selected set of mountain glaciers at the southern end of the Cordillera Blanca between 1984 and 2015 using satellite images. Furthermore, we analysed the existing glacier records in the Cordillera Blanca since the last glacial maximum to understand the evolution of glaciers in this region over a few centuries. There was a rise in the snowline altitude of glaciers in this region since the late 1990s with a few small glacier advances. Historical to the present El Niño-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) records were also analysed to understand whether there was a teleconnection between the glacier fluctuations in the region and the phase changes of ENSO and PDO. We also assessed the variations in three important climatic parameters that influence the glacier retreat—temperature, precipitation, and relative humidity—over a few decades. We calculated the anomalies as well as the seasonal changes in these variables since the mid-twentieth century. There was an increase in temperature during this period, and the decrease in precipitation was not so prominent compared with the temperature rise. There was an exceptionally higher increase in relative humidity since the early 2000s, which is relatively higher than that expected due to the observed rate of warming, and this increase in humidity is believed to be the reason behind the unprecedented rise in the snowline altitudes since the beginning of the twenty-first century.

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1 Introduction

Glacier retreat is normally associated with either a warming climate or a reduced rate of precipitation, and mountain glaciers are extremely sensitive to environmental fluctuations, particularly in the tropical region. Due to the specific climate conditions in the tropical region, ablation occurs at the lower part of the glaciers throughout the year and the response to the perturbations in climate, such as a rise in the air temperature, is rapid at the terminus (Rabatel et al. 2013). Major climate parameters that influence the mass balance of glaciers in this region are temperature, precipitation, solar energy (Silverio and Jaquet 2005; Burns and Nolin 2014; Alarcón et al. 2015), and humidity (Veettil et al. 2014), and hence, these tropical glaciers are good indicators of climate change.

The estimated rate of warming in the tropical Andes, which is about 0.11 to 0.15 °C/decade between 1939 and 2006 (Vuille et al. 2008a, 2008b; Bradley et al. 2009), is found to be two times the rate of global warming (0.06 °C/decade) during the same period (Bradley et al. 2006). This increased warming in the tropical Andes is hypothesised by many researchers (Hastenrath 1994; Vuille et al. 2008a, 2008b; Kaser and Osmaston 2002; Veettil et al. 2014, 2015a, 2015b) as one of the potential causes of enhanced glacier retreat occurring in this region which accommodates more than 99 % of all the tropical glaciers on the earth.

About 70 % of the earth's tropical glaciers are situated in Peru, and the Peruvian Cordillera Blanca is the most

Published online: 28 March 2016

extensively glaciated tropical mountain range in the world. The climate in Peru is highly influenced by the mountain barriers (the Andes). The Cordillera Blanca alone hosts 721 km^2 of glaciers (in 1970), which is about 35 % of the total Peruvian glaciers (Hidrandina, 1988) and is equal to 26 % of the tropical glaciers in the world (Kaser and Osmaston 2002). These glaciers function as freshwater reservoirs, particularly when the precipitation is sparse, due to seasonal melting (Silverio and Jaquet 2005). In fact, many of the catchments near the Cordillera Blanca are glacially fed (Baraer et al. 2015) and the discharge from these catchments is used for agriculture, mining, and hydropower and for human consumption (Vuille et al. 2008a). Based on a study in the Rio Santa watershed, Baraer et al. (2012) observed that the retreat of glaciers in the Cordillera Blanca is influencing the water availability near this catchment area during the dry season. Juen et al. (2007) also had similar results from the Llanganuco catchment in the same region.

The Cordillera Blanca was studied extensively by many researchers (Ames 1998; Silverio and Jaquet 2005; Solomina et al. 2007; Racoviteanu et al. 2008; Schauwecker et al. 2014) over a range of time periods, as this mountain chain has undergone several advances and retreats in glaciation. Like many other tropical Andean glaciers, the Cordillera Blanca lost a significant glacier mass after the Little Ice Age (Vuille et al. 2008a) and a recent study (Burns and Nolin 2014) showed that this mountain range has lost 25 % of its glacier area since the late 1980s. Based on lichenometry analysis, Solomina et al. (2007) observed major advances in the 1920s and smaller advances in the early 1970s (Ames 1998) with enhanced retreat between 1930 and 1940 and again an accelerated retreat since the late 1970s. Interestingly, Vuille et al. (2008a) observed a warming of 0.1 °C in the Cordillera Blanca between 1939 and 2006 which is well in phase with the higher glacier retreat.

Ocean-atmospheric phenomena such as the El Niño-Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO) influence the tropical Andean climate (Vuille et al. 2000; Vuille et al. 2008a, 2008b; Garreaud et al. 2009) and hence the glaciers in this region as well (Arnaud et al. 2001; Vuille et al. 2008a; Sagredo and Lowell 2012; Veettil et al. 2014, 2015a, 2015b). The pattern of the Southern Oscillation in South America and the observed patterns of precipitation, temperature, and humidity are highly correlated (Aceituno 1988). Opposite rainfall and temperature anomalies were observed during El Niño and La Niña events in the tropical Andes (Garreaud et al. 2009). El Niño events are normally associated with warmer than normal conditions and reduced quantities of precipitation. Cold (negative) and warm (positive) regimes of PDO, a climate index based on the North Pacific sea surface temperature (SST) variations, can persist for several decades and modulate the interannual relationship between ENSO and global climate (Mantua et al. 1997). There exists a combined influence of ENSO and PDO, particularly when these two are in phase, on tropical Andean glaciers (Veettil et al. 2014; Veettil et al. 2015a, 2015b), ENSO-related rainfall anomalies (Andreoli and Kayano 2005), and global land dry-wet changes in general (Wang et al. 2014).

The equilibrium line altitude (ELA) of a glacier separates the accumulation zone (annual mass balance >0) from the ablation zone (annual mass balance <0). The ELA is approximately equal to the annual maximum snowline during the dry season (austral winter) in the case of glaciers in the outer tropics (Rabatel et al. 2012). Due to the availability of Landsat series of images, it is possible to calculate the annual snowline variations since the 1970s. The main objectives of this study can be summarised as follows: (i) calculating the recent trends in snowline variations of a selected set of glaciers in the southern end of the Cordillera Blanca; (ii) assessing the variations in three key climatic parameters—temperature, relative humidity, and precipitation—that influence glacier mass balance in the study area; and (iii) evaluating the possibility of a teleconnection between the phase changes of ENSO and PDO and the trends in (i) and (ii).

2 Study region and climate conditions

Nevado Tuco (1 in Fig. 1) and Pastoruri glacier (2) (9° 55′ 45″ S; 77° 12′ 18″ W) are the southernmost snow-capped peaks (5479 and 5400 m, respectively) in the Cordillera Blanca second only to the Rahu Kutaq or Rajutuna (7) (5355 m). Nevado Tuco is also known as Tuku or Tucu. Nearby snow-covered peaks include Qiwllarahu or Queullaraju or Caullaraju (3), Jenhuararca (4), Huiscu or Challwa or Challhua (5), Condorjitanca (6), and Santun or Santon (8). Various glacial lakes nearby the Nevado Tuco are the source of Tuku River which flows towards the south. For understanding the cryosphere-climate interactions in the Cordillera Blanca, we observed the changes in the annual maximum snowline of selected glacier outlets from Nevado Tuco, Pastoruri Glacier, and a few nearby glaciated peaks including the south most Rajutuna. In order to understand the west-to-east as well as the elevation gradients in glacier response, we selected glaciers with varying altitudes from west to east. Qiwllarahu is on the western side whereas the Pastoruri (this one is classified as a Cirque glacier) lies on the eastern end. Nevado Tuco and Pastoruri are larger glaciers compared with Santun and Rajutuna, and the highest points of all these glaciers are situated above 5000 m a.s.l. The majority of glaciers in the Cordillera Blanca are classified as mountain-type due to their thin and steep nature (Burns and Nolin 2014).

There is a high seasonality in precipitation in this region, which is common in the outer tropics. Dry season occurs from May to September (austral winter), and snowfall occurs

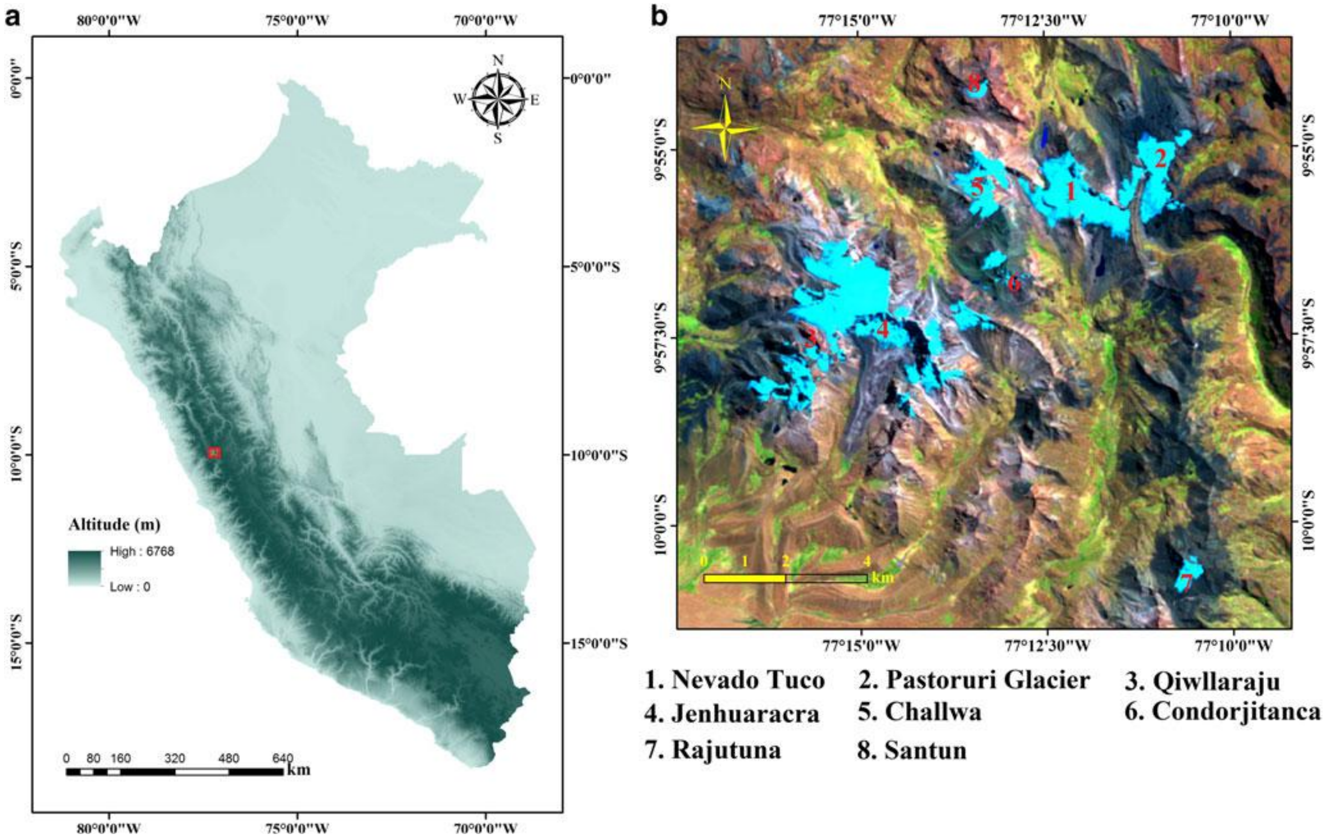

Fig. 1 **a** Relative location of the study site; **b** glaciers in the southern end of the Cordillera Blanca, including those considered in this study

during the wet season, normally from October to April. The dry and wet periods in this region were observed to be modified by the cold and warm regimes of ENSO (Garreaud et al. 2009), and the precipitation is governed by the movement of the Intertropical Convergence Zone (ITCZ) towards the Cordillera Blanca, thereby causing seasonal variations in humidity (Kaser et al. 2003). The observed increasing trend in humidity, which is important in partitioning the available energy for melting and sublimation, in the southern regions of Peru is moderate compared with that in the inner tropics (Ecuador and Colombia) and is higher when compared with that in the eastern cordilleras of Peru and Bolivia (Vuille et al. 2008a). Glacier melting is the main source of freshwater in catchments near the Cordillera Blanca during the dry season. Diurnal temperature variability is higher compared with the mean annual variability of air temperature.

Mass loss occurs throughout the year due to melting in summer and sublimation in winter, and accumulation occurs mainly during the warm and wet season (austral summer) at higher altitudes. Even though the glacier mass loss in the Cordillera Blanca due to sublimation is significant (Maussion et al. 2015), it is not a dominating mode as seen in the subtropics. Burns and Nolin (2014) reported that the glaciers in the northern parts on the Cordillera Blanca lost a smaller percentage of the glacier area compared with its southern counterpart. However, glaciers of comparable area and elevation must be considered before making such a conclusion, particularly when using remote sensing datasets.

3 Data and methodology

The datasets used for this study include medium-resolution satellite images, digital elevation models, meteorological data, and indices of ENSO and PDO. Landsat Thematic Mapper (TM), Enhanced Thematic Mapper+ (ETM+), and Operational Land Imager (OLI) images during 1984–2015 were used in this study, and all the OLI images were converted from 16- to 8-bits to apply the same algorithm. We used two Resourcesat-1 Linear Imaging Self Scanning Sensor (LISS-III) images in 2012 (23.5 m spatial resolution instead of 30 m of the Landsat images used) because of the unavailability of Landsat images in 2012. Radiometric and geometric corrections were provided to the Landsat images downloaded from the US Geological Survey (USGS) (http://landsat.usgs.gov/Landsat_Processing_Details.php). LISS-III images were geometrically corrected and georeferenced using the Landsat images. The images used for multitemporal analysis in this study were visually inspected and ensured the absence of cloud cover except in 1985. We used images acquired during (preferably towards the end of) the dry season (May–September). Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) Global Digital Elevation Model (GDEM) V2 that has a spatial resolution of 30 m is also used to calculate the snowline altitudes from the selected images. Images were processed using ERDAS IMAGINE 2013 and ArcMap 10 software packages.

Meteorological data such as temperature, precipitation, and humidity between 1950 and 2014 were obtained from multiple sources in the form of gridded datasets. For temperature, we used data from Climate Research Unit (CRU), University of East Anglia, and ERA-Interim data. Both these data have a latitude-longitude resolution of 0.5° × 0.5°. ERA-Interim data are available since January 1979 only, and CRU data used here were between 1950 and 2012. For relative humidity, we used National Centers for Environmental Prediction (NCEP)/National Centers for Atmospheric Research (NCAR) Reanalysis 1 data, which have a latitude-longitude resolution of 2.5° × 2.5°, and ERA-Interim data. We used CRU data for calculating the anomalies and seasonal variations in precipitation. One of the main drawbacks of the meteorological stations in the high-mountain environments is the difficulty of installation, and this may affect the accuracy of the datasets due to vertical gradients. It is seen that monthly precipitation varies within the western cordilleras from north to south in Peru. For example, the monthly precipitation near the Nevado Tuco is quantitatively higher compared with that near the Nevado Coropuna in the Cordillera Ampato, which is situated at a distance of 810 km towards south from the former one. The existing meteorological station near the study site is situated in Milpo at an altitude of 4400 m a.s.l. Even though incomplete or interrupted, we used some of the meteorological station data (Milpo) from the Servicio Nacional de Meteorología y Hidrología del Perú (SENAMHI, http://www.senamhi.gob.pe/) for comparison purposes.

The ENSO and PDO indices, from the National Oceanic and Atmospheric Administration (NOAA) (http://www.cpc.ncep.noaa.gov/) and the Japan Meteorological Agency (http://www.data.jma.go.jp/gmd/kaiyou/data/db/climate/pdo/pdo.txt), respectively, were also used in this research. We used Oceanic Niño Index (ONI), which is based on the SST anomalies in the Niño 3.4 region (5° N–5° S, 120°–170°W), for the time series analysis of the occurrence and phase changes of ENSO. The warm (El Niño) and the cold (La Niña) episodes were defined when a threshold of ±0.5 °C is met for at least five consecutive overlapping seasons. The threshold temperature value for defining the cold and warm regimes of PDO in the North Pacific (poleward of 20° N) is also ±0.5 °C. In this data, the monthly mean global average SST anomalies were removed to get rid of the signals of global warming from the PDO index (Zhang et al. 1997). Unlike ENSO, warm and cold regimes of PDO can persist for several decades.

As mentioned in the introduction, ELA separates the zone of accumulation from the ablation zone of a glacier and the fluctuation in ELA is a good proxy to study the climate change influence on the tropical glaciers. Since the areal changes of glaciers near our study site during the period 1975–2010 were already discussed in Alarcón et al. (2015), we analysed only the snowline changes in this research. Moreover, areal changes might not always be a good indicator of annual variations in climate due to the interannual variations in snow cover (Veettil et al. 2015a). For high-altitude mountain glaciers in the outer tropics, the maximum snowline altitude (SLA_{Max}) observed towards the end of the dry season (austral winter) can be used as an equivalent of the ELA and is pretty straightforward using remote sensing methods. This correlation between ELA and SLA_{Max} in the outer tropics was proved by Rabatel et al. (2012) by comparing the ELA of two glaciers (Artesonraju, Peru, and Glaciar Zongo, Bolivia) from field data with snowline altitude (SLA) calculated from satellite data. The satellite-based SLA_{Max} (the highest calculated SLA) during the dry austral winter was later used in many studies as equivalent to ELA in the tropical Andes (Veettil et al. 2014, 2015a, 2015b). In this study, we followed the method given in Rabatel et al. (2012) to estimate the SLA from Landsat images by creating a TM 5-4-2 false-colour composite image (6-5-3 for OLI images) after applying a suitable threshold value to near-infrared (TM 4) and green (TM 2) channels. The threshold digital number value applied to the near-infrared channel varies between 60 and 135 and that for the green channel between 80 and 160. The suitability of a particular threshold value depends on the lighting conditions (Rabatel et al. 2012), image acquisition date and time, and the presence of light cloud cover. True-colour image and 5-4-2 composite images before and after applying the threshold values (subset of a Landsat TM image acquired on 12 August 2008) are shown in Fig. 2, and as can be seen in the figure, the extent of the snow (light blue) is clearly separated from ice (black). Once the SLAs from various cloud-free images during the dry season are calculated, the highest observed SLA (SLA_{Max}) is selected as an equivalent of the ELA. For a detailed review of the spectral properties of snow and ice, methods for mapping them, and about how to select digital number threshold values for various wavebands from different spaceborne sensors, readers are requested to refer to Meier (1980) and Dietz et al. (2012).

Uncertainties in the calculated SLA (and hence SLA_{Max}) depend on the spatial resolution of the multispectral image used (30 m in this case), slope of the glacier surface, and/or

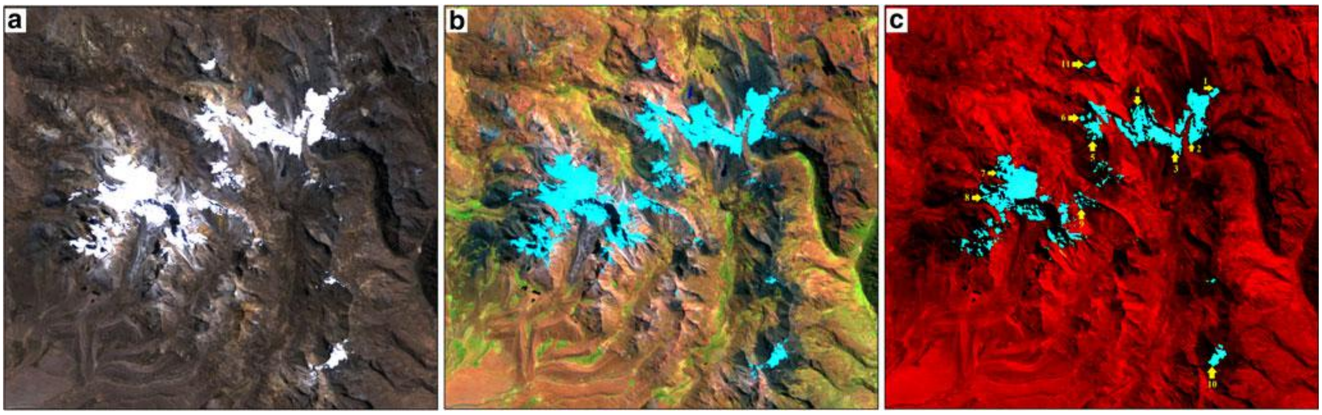

Fig. 2 Calculating the SLA of the study site. **a** True-colour TM image subset, **b** 5-4-2 false-colour composite of the same image subset before applying the threshold **c** after applying the threshold with selected outlet glaciers shown

the vertical accuracy of the DEM (Rabatel et al. 2012). Higher slopes at the upper parts of the glacier will result in higher inaccuracy in the estimated SLA, and hence, glaciers that have lower slopes were considered in this study (we used a slope of <20^{o} for all selected glacier outlets for convenience and avoided all steep terrain with the help of Google Earth 3D view). Since the vertical accuracy of ASTER GDEM V2 is about 20 m (at 95 % confidence) and this accuracy is confirmed to be in the same order of magnitude as the surface lowering of the selected glacier outlets, no altitudinal corrections were needed. We have not applied any accuracy measurement methods but considered unexpected snowfall events that may influence the accuracy of snowline estimation. Other than snowline variations during the period 1984–2015 using Landsat and LISS-III images, we also analysed a few glacier records that already exist in the Cordillera Blanca since the early fourteenth century to understand the evolution of glaciers before the existence of satellite images. In this record, we arranged whether there was a retreat or an advance during a particular period and compared it with the climate conditions prevailing at that time, particularly the phase changes of ENSO and PDO since the early fourteenth century.

We analysed the anomalies in precipitation, temperature, and humidity at the southern end of the Cordillera Blanca since 1950 using the gridded datasets from multiple resources described above. The gridded data were processed using MATLAB, and we averaged the four grids ([9.5° S; 77 W], [9.5° S; 77.5 W], [10° S; 77 W], [10° S; 77.5 W]) near the Milpo meteorological station (9° 55′ 27″ S; 77° 11′ 57″ W) for calculating the time series on a monthly basis and anomalies in the corresponding parameters. However, for NCEP/NCAR Reanalysis 1 data, we used only one grid for calculation because of its low spatial resolution. These meteorological data were also analysed for seasonal variations from 1950 to the present. The ENSO and PDO indices are already available up to date, and we created the time series directly using the data from NOAA and Japan Meteorological Agency, respectively (Fig. 3).

4 Results and observations

4.1 Glacier changes in the southern Cordillera Blanca during 1984–2015

Before calculating the recent snowline variations, we assessed the historical changes in the glacial behaviour in the Cordillera Blanca since the early fourteenth century from various papers published so far. The existing records of glacier retreat/advance in the Cordillera Blanca including the present study, the available records/reconstruction of the phase changes of PDO and ENSO, and the main references are summarised in Table 1. A brief discussion on the interrelationships among them is given in the next section. Some of the available studies were omitted because their study period cannot be used to link the phase changes of PDO and ENSO with glacier variations. A graphical representation of the results of snowline assessment of the selected glacier outlets from 1984 to 2015 is given in Fig. 4. For some glaciers, such as the Pastoruri glacier (snowlines 1 and 2) and Nevado Tuco (snowlines 3 and 4), we calculated snowlines on more than one side (see Fig. 2c). We had to exclude a few snowline calculations in 1985 due to the lack of completely cloud-free images. Interestingly, most of the glacier outlets with their initial snowlines situated above 4925 m have undergone a relatively slow rise during the study period. This relative stability of snowlines at higher altitudes was observed previously in southern Peru (Veettil et al. 2015b) and in Bolivian cordilleras (Veettil et al. 2015a). There was a slight decline in SLA_{Max} towards the end of the twentieth century and later continued to rise around the mid-2000s. It is observed that the snowline rise was rapid on higher slopes (snowlines 7 and 9) in comparison with lower slopes. A few glaciers on the north have shown relatively slower rise in

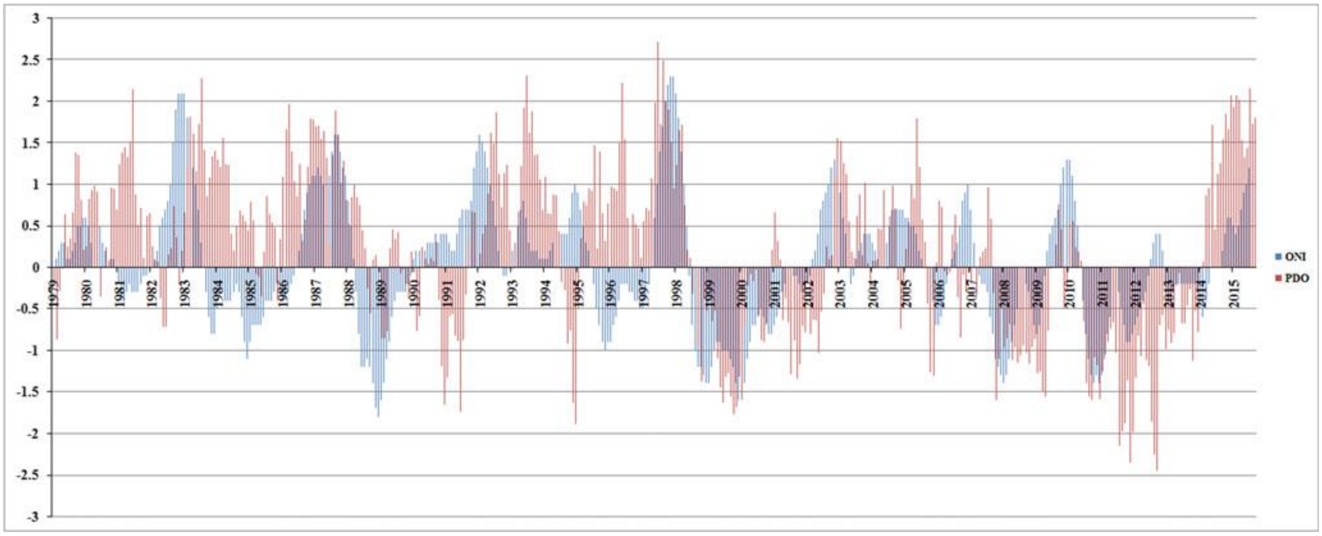

Fig. 3 ENSO (ONI) and PDO indices from January 1979 to September 2015

snowline during the study period (snowlines 6 and 11), even though dependent on their elevation, size, and slope. However, the distance between northern and southern glaciers in our study area is too short to generalise this observation.

Table 1 Summary of glacier fluctuations in the Cordillera Blanca, Peru, since the early fourteenth century and phase changes of ENSO and PDO.

4.2 Observing the climate trends near the study site

We calculated the variations in three key climatological parameters that influence the energy and mass balance of glaciers at the study site—temperature, precipitation, and relative humidity—during the recent six decades. Time series analysis provides a good tool for interpreting the possible causes of snowline changes of the selected glaciers.

4.2.1 Temperature

Time series of the monthly air temperature using ERA-Interim data between 1979 and 2014, from the beginning of the so-called Pacific shift (Ebbesmeyer et al. 1991) or the long-term PDO regime change, showed that there was definitely an increasing trend in temperature (0.42 °C/decade) during this period (Fig. 5). Both the ERA-Interim data and CRU data showed a rapid drop in temperature in 1998/1999 and then a quick recovering of the warming trend since the beginning of the twenty-first century. Similar trends were observed when we analysed the temperature for different seasons of the year (Fig. 6). The observed increase in temperature is higher compared with the temperature increase in the eastern cordillera in Bolivia (Veettil et al. 2015a) and the western cordilleras of Bolivia and southern Peru (Veettil et al. 2015b). We hypothesise that this increase is contributed partially by ENSO and PDO phase changes, at least before the strong El Niño event in 1997/1998. It can be seen from Fig. 5 that the observed trends in temperature are consistent with the observed phase changes of PDO and the warming trend is slightly weaker or even negative during some years after the strong El Niño event in 1997/1998, particularly when PDO entered its negative regime.

4.2.2 Humidity

Time series anomaly of the surface relative humidity showed a significant increasing trend, particularly after 2000, for the 1948–2015 period near Milpo (Fig. 7), and this observation is similar to that observed by Vuille et al. (2003) in the entire tropical Andes and by Salzmann et al. (2013) near the Cordillera Vilcanota region (which is situated in the southern wet outer tropics). Changes in the relative humidity for different seasons of the year are shown in Fig. 8. An increase in relative humidity in the atmosphere, particularly in the near surface levels, may cause a reduction in sublimation and an increase in melting due to a reduction in the vapour pressure difference between snow and air (Wagnon et al. 1999). This is how the rate of melting is higher during the austral summer (humid conditions) when the surface energy is used for melting instead of sublimation. An increase in humidity can contribute to an enhanced mass loss due to melting in the sublimation-dominated subtropics and outer tropics (mass loss due to sublimation increases from the inner tropics to the outer tropics and subtropics). It is proposed that this enhanced humidity influenced our study region at a lower rate, even though the observed increase in humidity is higher at the Cordillera Blanca, compared with the southern wet outer tropics (i.e. the eastern cordilleras of Peru and Bolivia). At the same time, the observed snowline reduction was higher compared with those glaciers in the dry outer tropics (western cordilleras of southern Peru, Bolivia, and northern Chile) for

Table 1 Summary of existing glacier fluctuations in the Cordillera Blanca, Peru, since the early fourteenth century and the phase changes of ENSO and PDO

Study period	PDO phase	El Niño/La Niña	References	Observations/comments
1330 ± 29	Cold phase	Transition period between Medieval Warm Period and Little Ice Age with a complex ENSO history	MacDonald and Case 2005 Solomina et al. 2007 Khider et al. 2011 Yan et al. 2011	Earlier glacial advances based on lichenometry
1630–1680	Cold phase after a long warm period	14 La Niña events during 1622–1639 with no El Niño in between, 6 Niño events during 1650–1661 then neutral until 1712	MacDonald and Case 2005 Solomina et al. 2007 Gergis and Fowler 2009 Jomelli et al. 2009	Maximum glacial extent observed
1730 ± 21	Slightly warmer than 1630–1680	7 El Niño events and 9 La Niña events during 1730–1743	Biondi et al. 2001 Vuille et al. 2008a Gergis and Fowler 2009	Recorded advances on several glaciers
1760 ± 19	,	4 El Niño events and 9 La Niña events during 1750–1770	Gedalof and Smith 2001 Vuille et al. 2008a Gergis and Fowler 2009	,
Mid-1920s	End of a prolonged cold phase	La Niña events during 1921–1926	Ames 1998; Kaser 1999	Reported short-lived advances
1930–1950	Warm phase	Two El Niño events were strong and continuous with short and weaker La Niña events	Kaser and Georges 1997; Georges 2004	Higher retreat with significant rise of ELA
1950–early 1970s	Cold phase	Both El Niño and La Niña occurred with stronger La Niña events until 1975	Hastenrath and Ames 1995	Comparatively slower retreat, reported on Yanamarey Glacier and a few advances
Mid-1970s	Cold phase		Kaser and Georges 1997 Kaser 1999	Small advance or slower retreat
Late 1970s towards the end of the twentieth century[a]	Beginning and end of the recent prolonged warm phase of PDO	Dominated by stronger and continuous El Niño events, including the strongest of the century and a strong La Niña in 1999–2000	Mark and Seltzer 2005 Silverio and Jaquet 2005 Racoviteanu et al. 2008	30–35 % decrease in the glacier surface area during 1962–1999, 15 % decrease during 1970–1996, 22.3 % decrease during 1970–2003
End of the twentieth century	Short-lived cold phase in late 1998 for 4 years	Relatively stronger La Niña events	Georges 2004	Very small advance
1984–2015	1984–1998 warm period and then the phases were broken down	Stronger and weaker El Niño and La Niña events with a prolonged and weaker El Niño during 1991–1995	This study	Slight increase in snowline during 1984–1995, near stability around early 2000s then rapid increase, particularly after 2010

[a] The twentieth century glacier retreat during 1977–2003 is found to be four times the rate observed during 1948–1977 (Mark et al. 2005)

some reason. This observation is without considering the elevation differences or atmospheric circulation patterns from the Amazon basin, and further investigation is needed here.

4.2.3 Precipitation

The overall linear tendency in precipitation is slightly decreasing between 1950 and 2012 as obtained from the CRU datasets (Fig. 9), and this result is in agreement with the observation by Vuille et al. (2003) in southern Peru between 1950 and 1994. The analysis of CRU datasets shows a slight decreasing tendency in seasonal precipitation, except during the austral summer (DJF) (Fig. 10), from the mid-1970s to the mid-1990s at the study site. At the same time, DJF precipitation from the mid-1970s to the mid-1990s shows a small overall increasing trend. It is also observed from the analysis of precipitation data for different seasons that the precipitation variations during the austral winter (JJA) were very small between 1950 and 2012.

As given in Fig. 3 and Table 1, the ENSO and PDO phase changes were considered for interpreting the possible causes of glacier changes due to variations in any of the parameters discussed above. In the next section, we discuss the glacier changes during the study period with focus on the phase changes of ENSO and PDO.

5 Discussion

Due to the continuous availability of Landsat data, a complete chronology of snowline changes during the recent three decades in the southern Cordillera Blanca is obtained from this study. The existing SLA (including SLA_{Max}) and ELA calculations in the Cordillera Blanca include Rabatel et al. (2012) on Artesonraju between 2000 and 2010, Dávila (2013) on Artesonraju for 2011–2012, and Gurgiser et al. (2013) on Shallap Glacier for 2006–2008. Many other studies calculated the areal changes (Silverio and Jaquet 2005; Racoviteanu et al.

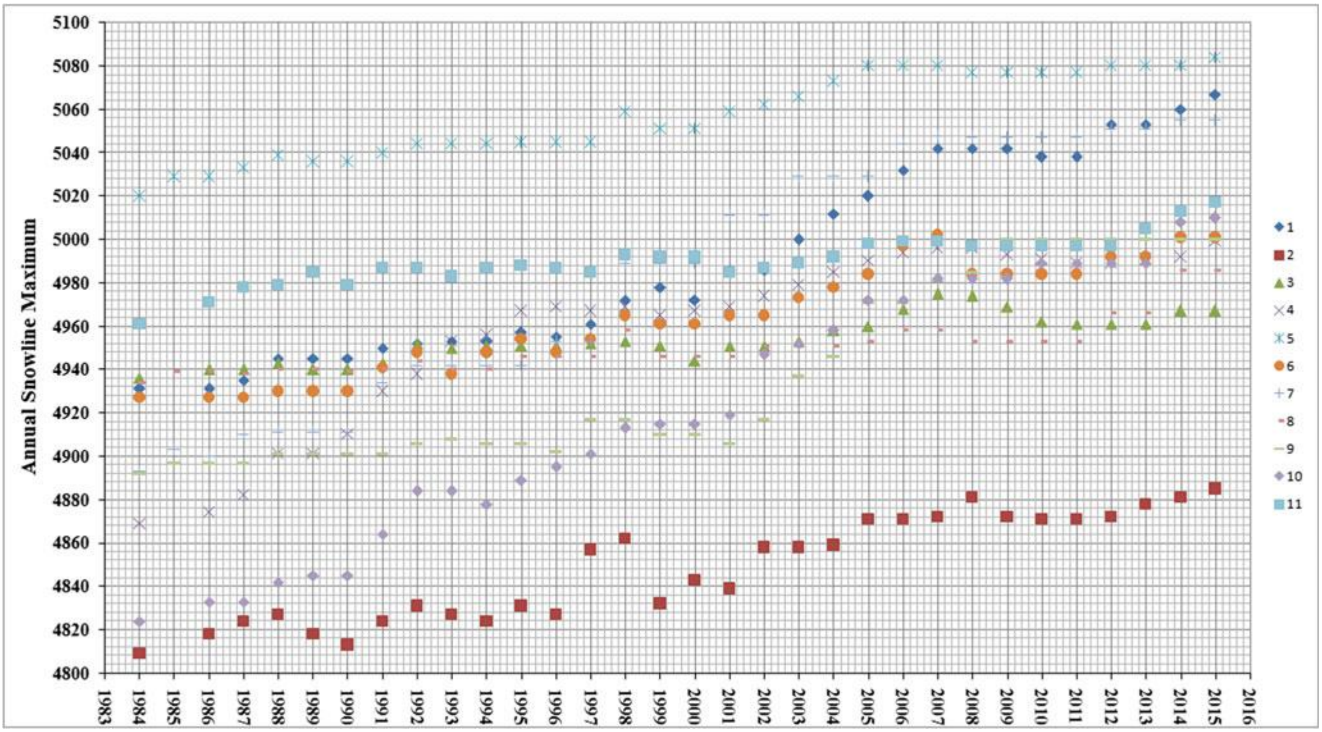

Fig. 4 Variations in SLA_{Max} in the southern Cordillera Blanca from 1984 to 2015. Each *curve* represents the variations in the SLA_{Max} of corresponding glacier outlet in Fig. 2c

2008; Vuille et al. 2008b; Burns and Nolin 2014) in the Cordillera Blanca, and our results on snowline variations are comparable with the results from these studies. About 25 % of the glacier area was lost between 1987 (643.5 km^2) and 2010 (482 km^2), and glaciers situated towards the south lost a greater percentage compared with their northern counterparts in the Cordillera Blanca (Burns and Nolin 2014). Racoviteanu et al. (2008) calculated a decrease of 22.4 % in the glaciated area between 1970 (Ames et al. 1989, 723.3 km^2) and 2003 (569.6 km^2) in the same study area. The higher loss of glacier area in the south (Burns and Nolin 2014) can be due to lower altitude or due to the proximity of the January ITCZ as hypothesised by Veettil et al. (in review). Recently, Alarcón et al. (2015) also confirmed the rapid retreat of glaciers at the southern end of the Cordillera Blanca (our study site) and an area loss of about 58 % was observed between 1975 and 2010. Minute discrepancies in quantifying the area changes (and hence mass balance changes too) may arise from the differences in regional climatic factors in comparison with results from a global-scale research and field work data and even from the differences in spatial resolution of the images used.

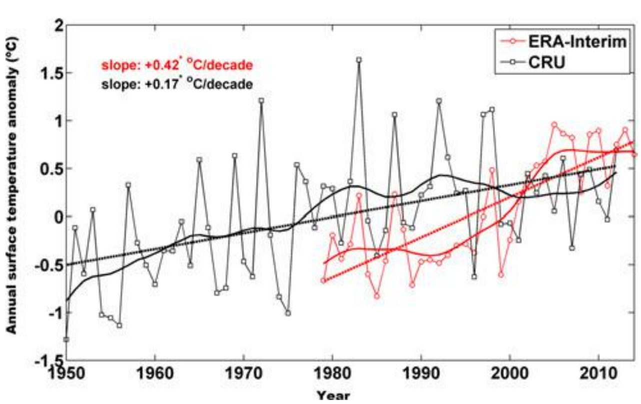

Fig. 5 Time series of annual surface temperature anomaly at 500 hPa over Milpo using ERA-Interim datasets (*red*) from 1979 to 2014 and CRU datasets (*black*) from 1950 to 2012 based on the period 1961–1990. *Thick curves* denote the 11-year Gaussian-type filtered values, and the *dashed lines* are the linear trends. One *asterisk* indicates that a trend exceeds the 95 % confidence level. Here, the anomaly is calculated based on the period 1981–2010 for ERA-Interim and 1961–1990 for CRU

The key climatological variables that influence the glacier evolution are temperature, precipitation, humidity, cloudiness, and solar irradiance. In a new framework (Sagredo and Lowell 2012; Sagredo et al. 2014) to classify the Andean glaciers based on the meteorological conditions (temperature, precipitation, and humidity), glaciers in the Cordillera Blanca belong to the northern wet outer tropics. In this region, air temperature is seasonally uniform (near 0 °C) and the average humidity is about 71 %. Schauwecker et al. (2014) observed a higher warming of 0.31 °C/decade between 1969 and 1998 whereas the warming slowed down to 0.13 °C/decade between 1983 and 2012 using data obtained from a large number of meteorological stations near the Cordillera Blanca as well as NCEP/NCAR- and ERA-Interim-gridded datasets. Our observations are in agreement with this study. The same study (Schauwecker et al. 2014) documented a decrease in

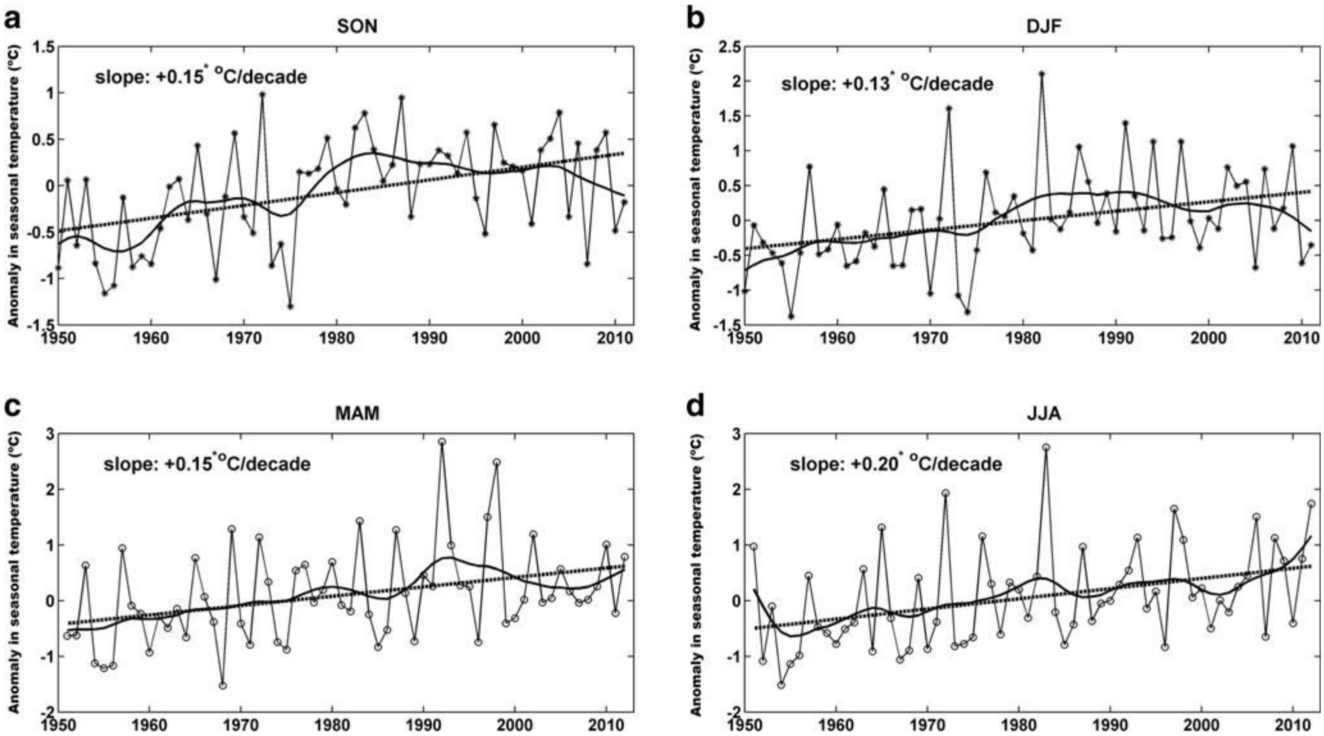

Fig. 6 The same as Fig. 5, but for different seasons, from CRU datasets

maximum daily air temperatures as well as an increase in precipitation of nearly 60 mm/decade since the early 1980s. In our study, a small decrease in precipitation is observed near Milpo, particularly during JJA, over the last six decades and a slight increase in DJF precipitation between 2000 and 2010. The observed decrease in precipitation, however, is not that sufficient to cause an enhanced glacier retreat in this region. An increase in DJF precipitation along with a rise in the air temperature can contribute to enhanced glacier retreat because rainfall dominates instead of snow under warming conditions which in turn causes a decrease in surface albedo. However, it should be noted that rainfall rarely occurs at higher elevations in the outer tropics, particularly near the subtropics. Other studies such as Mark and Seltzer (2005) also observed a strong warming over the Cordillera Blanca between 1962 and 1999. If the warming causes a rise in the freezing altitude, this can lead to higher melting because of rain instead of snow at the glacier margins (Bradley et al. 2006). The results from CRU data used in this study are in agreement with Mark and Seltzer (2005). The causes of exceptional increase in humidity, from

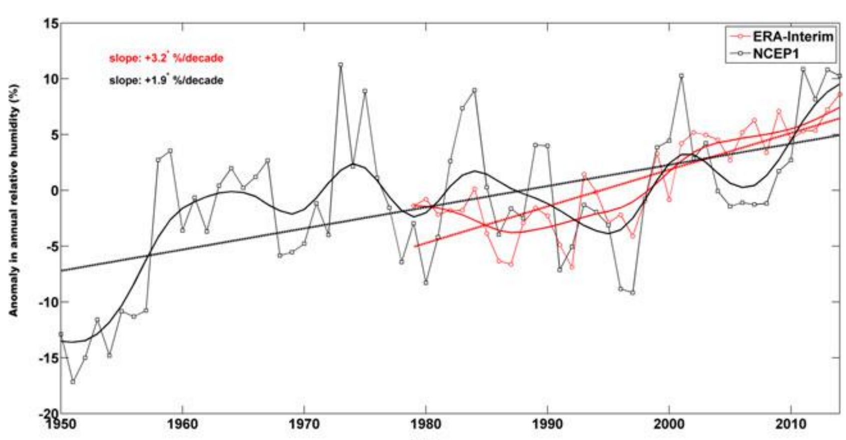

Fig. 7 Time series of annual mean relative humidity (RH) anomaly at 500 hPa near Milpo from ERA-Interim datasets (*red*) for the period 1974–2014 and from NCEP/NCAR Reanalysis 1 datasets (*black*) for 1948–2014. *Thick curves* denote the 11-year Gaussian-type filtered values, and the *dashed lines* are the linear trends. One *asterisk* indicates that a trend exceeds the 95 % confidence level. Here, the anomaly is calculated based on the period 1971–2010 for ERA-Interim and 1961–1990 for NCEP/NCAR Reanalysis 1

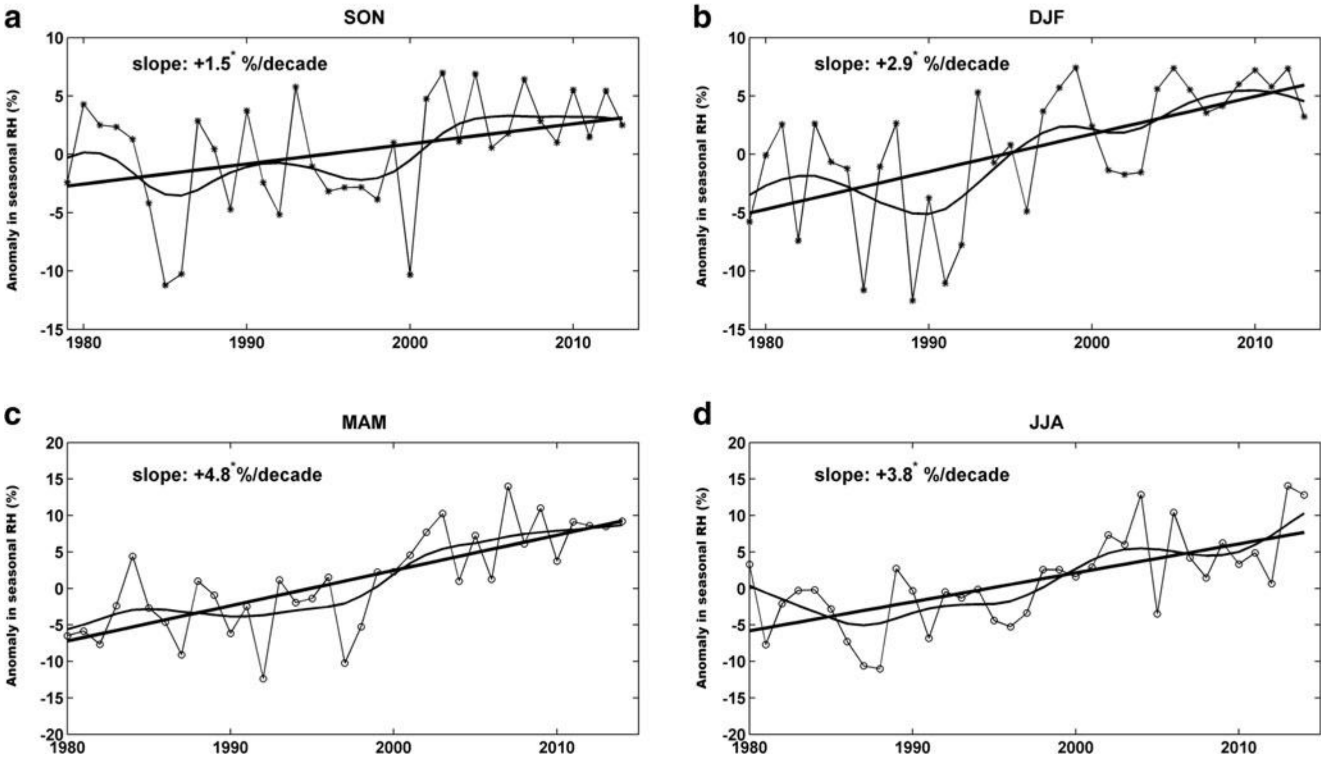

Fig. 8 The same as Fig. 7, but for different seasons, from ERA-Interim datasets for 1979–2014

our study, since the early 2000s is still not understood well because there was a small negative temperature anomaly between 2000 and 2008 (Fig. 5 and Fig. 7). It is seen that the observed increase in relative humidity, and therefore the specific humidity, is higher than that expected from the variations in temperature (between 2000 and 2008). The enhanced humidity, instead of temperature changes, is believed to be the cause of exceptional snowline rise in this region due to the change from sublimation-dominated to melt-dominated mass loss. The humidity increase before the late 1990s, however, was consistent with the observed warming (Vuille et al. 2003; Mark and Seltzer 2005). In comparison with another study (Veettil et al. 2015a) on snowline variations in the dry outer tropics, it is observed that the glaciers in the Cordillera Blanca are under a higher threat of rapid retreat, particularly after 2010. This observation fits into the exceptional increase in humidity rather than increase in temperature or decrease in precipitation. Moreover, glaciers in the outer tropics are relatively slow in response to temperature changes in general compared with those glaciers in the inner tropics (Favier et al. 2004).

Recent studies suggested higher rates of warming at higher altitudes (Bradley et al. 2006; Rangwala 2013; Pepin et al. 2015), the so-called elevation-dependent warming (EDW). The factors behind this phenomenon are recognised as complex snow-albedo mechanism, cloud cover, higher humidity,

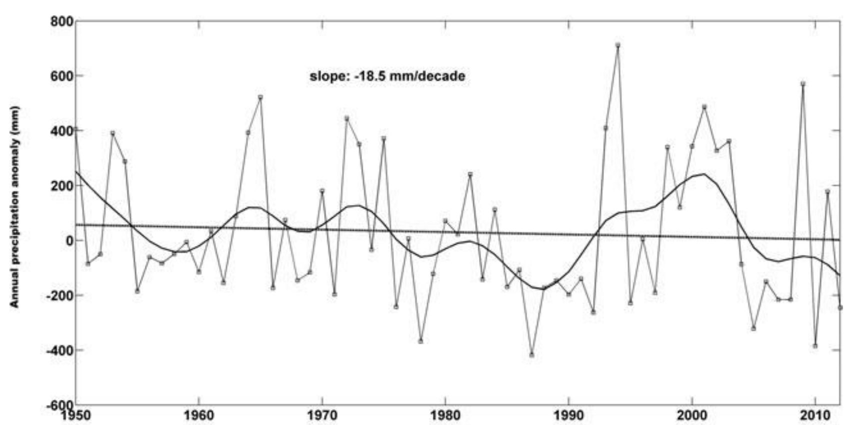

Fig. 9 Time series of annual precipitation anomaly (*thin black curves with dots*) near Milpo from CRU datasets for the period 1950–2012. The *black curve* is 11-year Gaussian-type filtered values, and the *dashed black line* illustrates the linear trend. Here, the anomaly is calculated based on the period 1961–1990 on a decadal scale

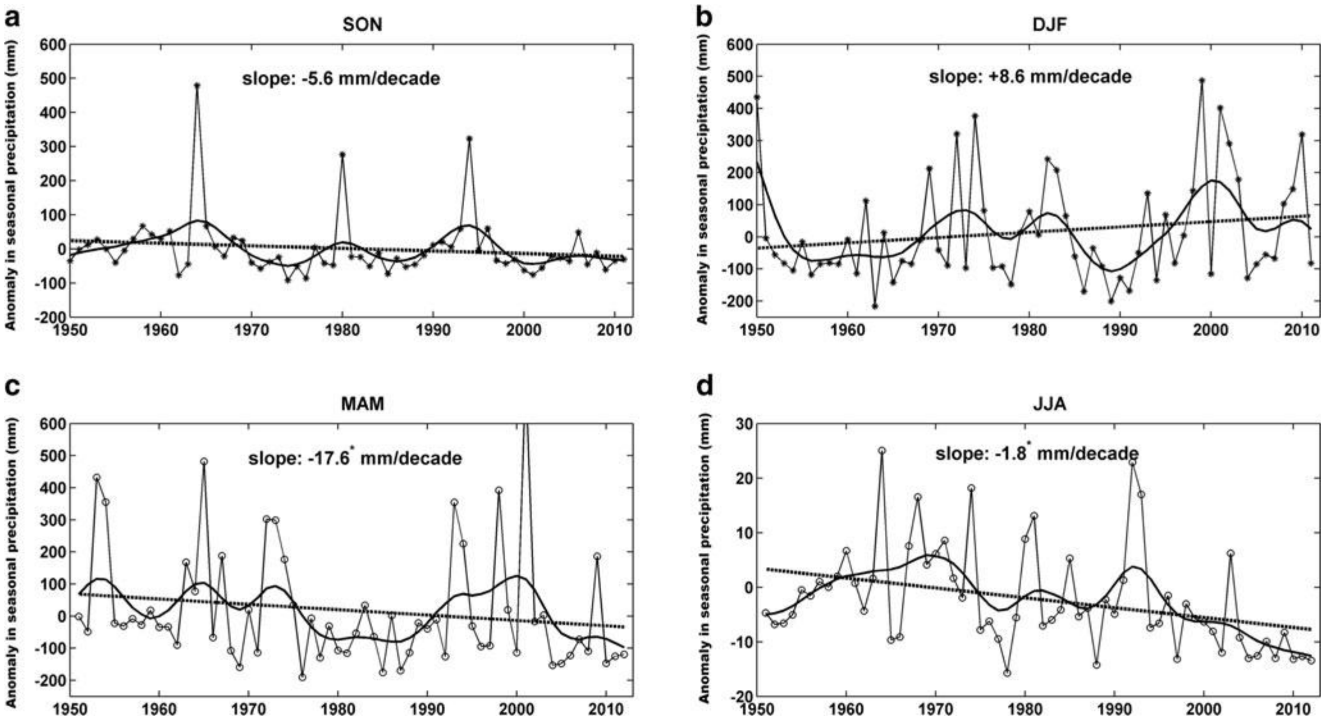

Fig. 10 The same as Fig. 9, but for different seasons, from CRU datasets

atmospheric aerosols, or a combination of all these together (Pepin et al. 2015). Our results are important in this context because a recent hiatus in global warming (Vuille et al. 2015) is observed, particularly along the Pacific coast. However, warming still continues at higher elevations of the tropical Andes, which is evident from the snowline variations. Even though higher humidity is one of the reasons (Rangwala 2013) behind EDW, further investigation is needed to understand whether some other factors also contribute to the observed snowline variations in the Cordillera Blanca. Furthermore, Pepin et al. (2015) hypothesised that the interannual to decadal variability in large-scale circulation such as ENSO and PDO can influence regional variability in EDW, particularly in tropical mountains (Diaz et al. 2014).

The shift in PDO from its prolonged negative regime between 1947 and 1976 to its positive regime in the late 1970s might have contributed to the observed warming from 1979 to 1998 in the tropical Andes (Garreaud et al. 2009). Moreover, the strongest El Niño in 1997/1998 along with many other El Niño episodes between 1991 and 1995 contributed to these increased warming conditions. A strong correlation between ENSO and surface energy balance and surface mass balance (SEB/SMB) was already found in the Cordillera Blanca based on a study on the Shallap Glacier (Maussion et al. 2015), and this correlation was quantitatively higher compared with previous studies (Vuille et al. 2008b). The geographical position of the Cordillera Blanca (with respect to the Niño 3.4 region) would definitely influence this increased correlation between the glacier changes and ENSO. Moreover, being relatively far from the January ITCZ (Veettil et al., in review) and being less influenced by the Amazon and Atlantic circulation patterns (Sagredo and Lowell 2012; Sagredo et al. 2014), Pacific influence determines the majority of the climate in the Cordillera Blanca. It is known in the tropical Andes that when PDO occurs in phase with ENSO, the influence of ENSO on glaciers is stronger and more visible compared with when ENSO is in opposite phase to PDO or occurs during PDO neutral periods (Veettil et al. 2014, 2015a, 2015b). A weak correlation was observed between the annual precipitation variability and PDO in the Cordillera Blanca during the recent years (Schauwecker et al. 2014), and this could influence the mass balance changes as well (Veettil et al. 2015a, 2015b). However, a strong correlation occurred between air temperature and PDO before about 1995 (Schauwecker et al. 2014) after the PDO regime shift in 1976. It is interesting to note that a weak/no correlation between PDO and temperature occurred exactly when the PDO decadal cycles were broken down, more or less in the late 1990s (the warm regime of PDO occurred during 1977–1998, entered a 4-year negative phase during 1998–2001, 3 years from 2002 to 2005 warm phase, neutral until 2007, 6 years cold phase during 2008–2013, and now started a warm phase). The signals of this phenomenon, even though not uniform, are visible from the snowline variations in the Cordillera Blanca too.

Without knowing the glacier response to climate change and decadal/annual climate variability in the past, the present knowledge on the glacier-climate interactions would be incomplete. We do not have mass balance or ELA values derived from remote sensing datasets before the early 1970s. Fortunately, historical ENSO and PDO data are available from around the beginning of the twentieth century and many field data are available on the glacier fluctuations in the Cordillera Blanca (Kaser et al. 1996; Kaser and Georges 1997; Ames 1998; Kaser 1999; Georges 2004; Mark and Seltzer 2005; Mark et al. 2005; Raup et al. 2006; Jomelli et al. 2009). There are a few reconstructed ENSO and PDO indices available for a few centuries (Biondi et al. 2001; Gedalof and Smith 2001; MacDonald and Case 2005; Gergis and Fowler 2009; Khider et al. 2011; Yan et al. 2011). In fact, as hypothesised by Jomelli et al. (2009) based on the historical data analysis of ENSO events in Peru by Ortlieb (2000), a clear shift in the frequency of El Niño events since the seventeenth century could have triggered the end of the Little Ice Age in the Andes. The *maximum* glacial extent between 1630 and 1680 (Solomina et al. 2007) occurred during a cold regime of PDO with many La Niña events together with a few or no Niño phases in between (Table 1). However, low accuracy of reconstructed ENSO and PDO indices may add some uncertainties to the results. Glacier advances and retreats in the Cordillera Blanca observed by Kaser (1999) overlap with the phase changes of PDO since the early 1920s. The observed glacier advance in the mid-1920s (Ames 1998; Kaser 1999) coincides with a short-lived cold regime of PDO with prolonged La Niña events during 1920–1925, and there was a rapid retreat from 1930 to 1950. Surprisingly, the period between 1925 and 1946 was witnessing a prolonged warm regime of PDO, even though the frequency of El Niño and La Niña events was balanced. Similar (advancing) trends were observed during the early 1970s (Kaser and Georges 1997), where PDO negative phase prevailed from 1947 to 1976, followed by a rapid glacier retreat between the late 1970s to the mid-1990s (Mark and Seltzer 2005). The frequency of El Niño events has increased since the late 1970s. A small advance was observed at the end of the twentieth century by Georges (2004), and strikingly, this advance coincides with the short-term 4-year negative PDO from 1998 to 2001. Comparable trends in glacier changes were observed in the Cordillera Real in Bolivia (Rabatel et al. 2005, 2006). However, because the PDO cycles have broken down recently and local microclimate also influences glacier mass balance, it would be more complex to establish a correlation between glacier changes and the short-term PDO/ENSO phase changes in recent years.

Solar irradiance modulation is an external forcing that can influence the glacier fluctuations (Jomelli et al. 2009), which is not considered in many researches. An increase in the incoming solar radiation due to decrease in cloud cover can influence the sensitivity of glaciers towards a change in one or many of the parameters such as air temperature that influence the stability of mass balance (Rabatel et al. 2006). However, long-wave radiation can provide energy for melting on high-albedo surfaces in mountain environments during cloudy skies (Sicart et al. 2010) to some extent. In any case, the solar irradiance plays a definite role in the albedo reduction and glacier melting in the tropical mountain areas. The glacial maximum in Peru and Bolivia and the Maunder solar minimum (1645–1715) occurred concurrently and again during a less significant Dalton solar minimum (1783–1830), the glacier retreat in the tropics slowed down and even a few advances were reported (Rabatel et al. 2005). However, the coincidence of minimum solar irradiance and glacial advances in the tropics cannot be considered as a global phenomenon (Rabatel et al. 2005) and in fact, many glaciers in the Alps were reported to have retreated or not advanced during the Maunder minimum (Luterbacher et al. 2001). For example, despite the low temperatures during the Maunder minimum, no glacier advances were reported in the Swiss Alps (Pfister 1984). In fact, glacier advances in the early nineteenth century in Europe were more extensive than during the Maunder minimum (Zasadni 2007). There was an increase in solar irradiance since the early 1900s until the late 2000s, and currently, we are going through a period having a decreasing tendency since the late 2010s (Fröhlich 2000, 2003, 2006; Lean 2000). Updated solar irradiance data can be obtained from the World Radiation Center (ERC), Physikalisch-Meteorologisches Observatorium Davos (http://www.pmodwrc.ch/) for further research.

Last but not least, volcanic activities influence glacier mass balance changes both directly (Ginot et al. 2010) and indirectly (Rabatel et al. 2013, Veettil et al. 2015a, 2015b) and may be able to mask the influence of climate variations. Direct influence can be due to glaciochemical activities of ions or reduction in the surface albedo due to particulate materials. Indirect influence includes the reduction in the solar irradiance due to aerosols such as sulphates in the stratosphere. This type of indirect influence was observed to have occurred during the Mt. Pinatubo eruption in 1991, which masked the influence of a series of El Niño events between 1991 and 1995 (Veettil et al. 2015a, 2015b). In this study also, we came across a 'slower than expected' rise in the snowline during the El Niño events in the same period.

6 Conclusions

We quantified the changes in annual maximum snowlines (SLA_{Max}) of selected glaciers at the southern end of Cordillera Blanca and assessed the climate conditions between 1984 and 2015. Below are our main conclusions:

1. Glaciers in the Cordillera Blanca are undergoing a rapid retreat under the current climate conditions, but this retreat is relatively slower compared with the observed retreat in the inner tropics or the southern wet outer tropics (eastern cordillera) in Bolivia.
2. The phase changes of PDO coincide with the retreat and advance of glaciers in this region since the beginning of

the twentieth century. However, in the recent decades, particularly since the end of the twentieth century, the co-occurrence of positive (negative) PDO and retreat (advance or stability) of glaciers in this region is not so evident and we assume that this could be due to the broken down cycles of PDO.

3. A strong coincidence of El Niño events and glacier retreats was observed only if these events persist for a long time, if too strong (ONI above 1.5) or occur in phase with the positive regime of PDO. This can be altered by regional climate settings, volcanic activities as well as the amount of solar irradiance/cloud cover.
4. We could observe slightly different rates of snowline variations at the northern and southern sides of the study region as observed in other researches (Burns and Nolin 2014; Alarcón et al. 2015). However, the area considered in this study is too small to generalise such differences. Glaciers with lower elevation fluctuated at a higher rate compared with those at higher altitudes, and similar results were observed in the outer tropics (Veettil et al. 2015b).
5. We observed that the rate of snowline variations at our study site is smaller compared with that observed in the eastern cordilleras of Bolivia (Veettil et al. 2015a) and higher compared with the snowline variations in the dry outer tropics (Veettil et al. 2015a, 2015b).
6. The observed snowline rise might be partially caused by the warming conditions that induced rainfall instead of snowfall, particularly during DJF and partially by the increase in humidity that enhances melting instead of sublimation, particularly during the austral winter.

Acknowledgments First author acknowledges Fundação de Amparo à Pesquisa do Estado do Rio Grande do Sul (FAPERGS) for his research funding. We would like to express our sincere thanks to the US Geological Survey (USGS) for the Landsat images, Instituto Nacional de Pesquisas Espaciais (INPE) for Resourcesat images, the National Oceanic and Atmospheric Administration (NOAA) for the ENSO indices, and the Japan Meteorological Agency for the PDO indices. The University Corporation for Atmospheric Research (UCAR), Climate Research Unit (CRU) of the University of East Anglia, and NOAA are highly acknowledged for the meteorological datasets. We are indebted to three anonymous reviewers for their constructive comments.

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CAPITULO VI

Influence of ENSO and PDO on mountain glaciers in the outer tropics: case studies in Bolivia

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Theoretical and Applied Climatology, 125, 757-768, 2016

Theor Appl Climatol (2016) 125:757–768
DOI 10.1007/s00704-015-1545-4

ORIGINAL PAPER

Influence of ENSO and PDO on mountain glaciers in the outer tropics: case studies in Bolivia

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Received: 13 May 2015 / Accepted: 26 June 2015 / Published online: 7 July 2015

Abstract This paper emphasize on the observational investigation of an ice-covered volcano and two glaciated mountains in the Central Andes from 1984 to 2011. Annual snowlines of the Nevado Sajama in the Cordillera Occidental and the Nevado Cololo and the Nevado Huanacuni in the Cordillera Apolobamba in Bolivia were calculated using remote sensing data. Landsat TM, Landsat ETM+, and LISS-III images taken during the end of dry season were used in this study. Changes in the highest annual snowline during May–September is used an indirect measure of the changes in the equilibrium line altitude of the glaciers in the outer tropics. We tried to understand the combined influence of the El Niño-Southern Oscillation and the Pacific Decadal Oscillation on the variations in the annual snowline altitude of the selected glaciers. Meteorological data in the form of gridded datasets were used for calculating the anomalies in precipitation and temperature during the study period. It is found that the glaciated areas were fluctuated with the occurrence of warm and cold phase of ENSO but the magnitude of the influence of ENSO is observed to be controlled by the phase changes of PDO. Snowline of the Nevado Sajama fluctuated heavily when cold and warm phases of ENSO occur during the cold and warm regimes of PDO, respectively. Nevado Cololo and Nevado Huanacuni are showing a continuous retreating trend during the same period. This clearly indicates that the changes in the Pacific SST patterns have more influence on glaciers in the Cordillera Occidental compared with those in the Cordillera Oriental of the Bolivian Andes.

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1 Introduction

Global climate is highly complex and extends over a wide range of spatial scale with a large number of subsystems varying on different timescales. Tropical and subtropical glaciers are extremely sensitive to climate change, and their response to any small imbalances in the environment is relatively rapid (Arnaud et al. 2001). The Andes occupies about 99.7 % of all the tropical glaciers in the world (Kaser 1999). It is predicted that the tropical Andes might experience a massive warming in the order of 4.5–5 °C by the end of twenty-first century and an increase in precipitation during the wet season and a decrease in the same during the dry season (Vuille et al. 2008). Precipitation rates were found to be poorly correlated with rain-induced glacier ablation and runoff, where precipitation is in solid form (Francou et al. 1995). It is found that warming in the upper troposphere in the tropical region is greater than the global mean due to increased latent heat release, and this caused the recent increase in tropical precipitation (Mitchell et al. 1990). With no change in precipitation, an increase in temperature can lead to the disappearance of glacial coverage in the tropics (Rabatel et al. 2013). This can lead to increased shortage of freshwater for domestic and agricultural usage to a majority of population, particularly in Peru and Bolivia, due to the reduction in quantity of meltwater originating from glaciers.

The Andes in South America is passing through a large number of temperature and precipitation zones with higher influence of Atlantic circulation patterns in the north and Pacific influence in the south (Sagredo and Lowell 2012). Both circulation patterns influence the central Andean region. Easterly wind anomalies favor wet conditions in the central Andes whereas westerly wind enhances dry conditions (Vuille and Keimig 2004). Some of the glaciers in Bolivia, such as Glacier Charquini, lost about 65–78 % of its little ice age (LIA) surface area, and the equilibrium line have been increased by 160 m (Rabatel et al. 2006). Interannual variability and longer-term trends in mass balance changes of larger and smaller glaciers were found to be very similar (Vuille et al. 2008). Even though many climate models exist that predict the future changes of the tropical glaciers, many of them ignore the changes in the ocean (salinity, for example) and ocean circulations that contributes El Niño-Southern Oscillation (ENSO) and other decadal oscillations that influences the present distribution and frequency of tropical storms. ENSO events are associated with drought in the Altiplano, which favors glacier melting and retard runoff in the non-glaciated areas (Ribstein et al. 1995). ENSO events are characterized by a markedly negative mass balance with a water depletion equivalent to twice the amount of the accumulated precipitation, an elevation in the equilibrium line altitude (ELA), a reduction in the accumulation-area ratio (AAR), and a substantially reduced accumulation rate at high altitude. A decrease in the glacier accumulation with the warm phase of ENSO was confirmed based on the studies on the Quelccaya ice cap in Peru (Thompson et al. 1984) and also based on the analysis of water records from Glaciar Zongo in Bolivia (Ribstein et al. 1995). ENSO controls majority of the interannual variability of mass balance at a regional scale in the Cordillera Real (Rabatel et al. 2006). However, correlating these effects just with short-lived ENSO events does not seem to be promising. Recent studies suggest a combined effect of ENSO with a long-lived, ENSO like, Pacific Decadal Oscillation (PDO) on tropical climate (Kim et al. 2014; Veettil et al. 2014). Other southern hemisphere oscillations like Antarctic Oscillation (AAO), which is the counterpart of Arctic Oscillation (AO) in the northern hemisphere, could be considered while studying the influence of complex circulation anomalies in the tropical South America. Atlantic sea surface temperature (SST) variability cannot be considered as completely independent on the Pacific SST variability (Enfield and Mayer 1997), particularly during the boreal spring. By considering the combined effects of all these phenomena, we might be able to explain the influence of climate variations on tropical Andean glaciers that were not or least explained based only on the ENSO.

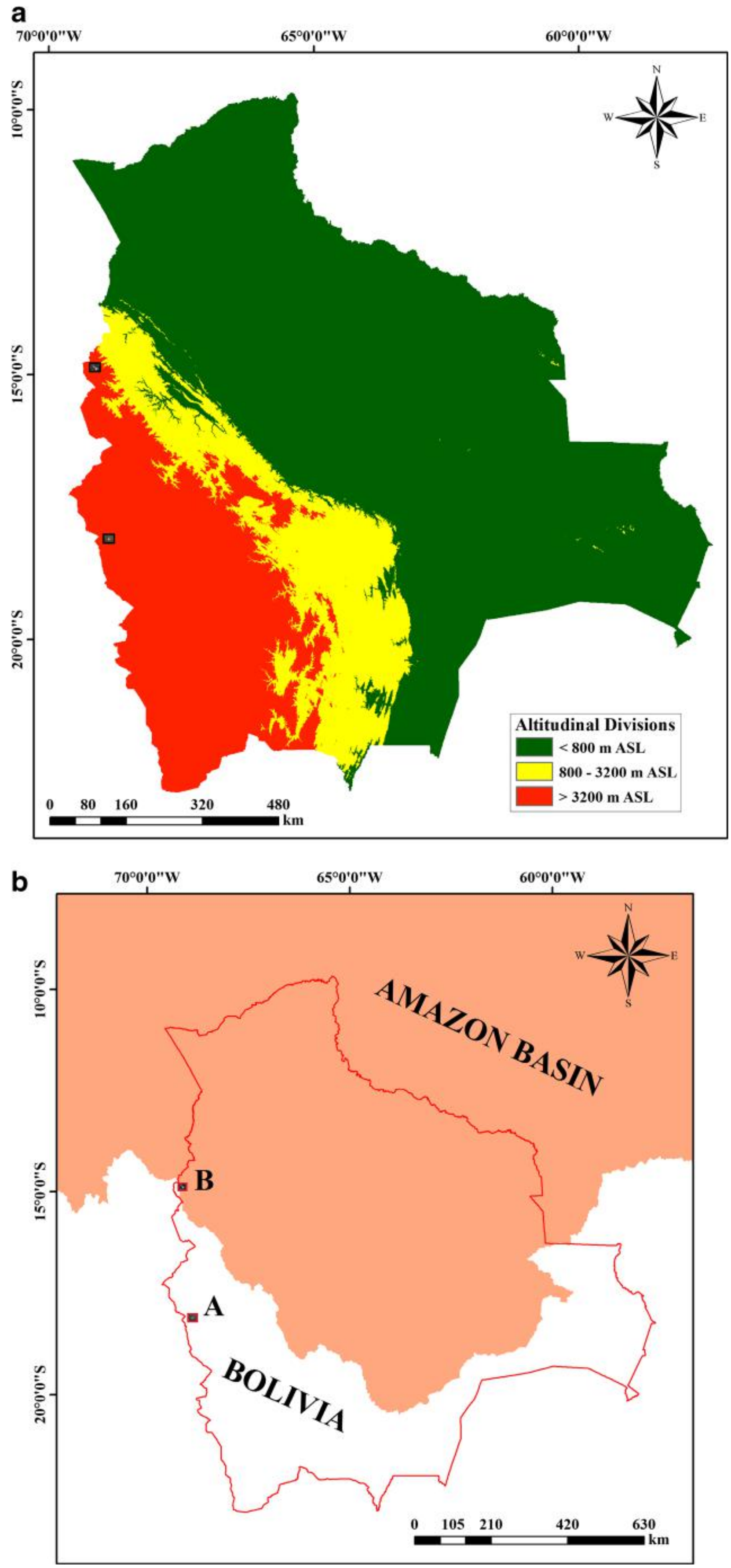
Even though sometimes limited by spatial, spectral, and temporal properties, recent advances in remote sensing and photogrammetric techniques helped the scientific community to understand recent surface and mass balance changes, and to study the evolution of tropical mountain glaciers, at least from the mid-twentieth century. Many researchers (Arnaud et al. 2001; Rabatel et al. 2012) used remotely sensed images and digital elevation models (DEM) to calculate the snowline altitude (SLA) that can be used to calculate an approximate value of the ELA in the outer tropics. This ELA can be used to calculate the approximate mass balance change, which in turn helps to understand the climate dynamics in the tropical Andes. In this study, we tried to explore the possibility of using the annual snowline maximum of glaciers in the western and eastern cordilleras of the Bolivian Andes as an equivalent of the ELA and attempted to correlate the snowline variations with the anomalies in monthly precipitation and temperature. We also tried to find if any teleconnection exists between the glacier variations and the occurrences of ENSO, PDO, and AAO.

2 Study sites and climate conditions

Bolivia is considered as a tropical country with its main altitudinal divisions consisting of the lowlands (<800 m above sea level (ASL)), the Andean slopes (800–3200 m ASL), and the highlands or the Altiplano (>3200–6500 m ASL) (Fig. 1a). Climate in Bolivia varies from tropical to cold desert climate depending on the altitude (Seiler et al. 2013a). Annual mean surface air temperatures vary from 0 to 30 °C, and precipitation ranges from less than 300 to 3000 mm per year. Precipitation in Bolivia and its interannual variability is linked to the tropical SST anomalies and atmospheric circulation patterns (Arnaud et al. 2001; Vuille, 1999). Majority of the precipitation occurs during December–March, and the austral summer (DJF) is characterized by a low-pressure system that enhances easterly trade winds to transport moisture from the Atlantic (northern tropics) to the continent. This moisture content, which is deflected by the Andes, is transported towards the south and causes enhanced precipitation in the Atlantic Ocean. Condensational heat release over the Amazon occurs simultaneously, and the Andean slopes cause the formation of an upper-level Bolivian high-pressure system, which causes enhanced transport of moisture from the Amazon to the Bolivian highlands and lowlands (Seiler et al. 2013a; Vuille 1999). In the austral winter (JJA), less moisture transport occurs from the northern tropical Atlantic to the continent and the cold fronts from the South Pole penetrate into the Bolivian lowlands thereby lowering the temperature and limit the precipitation (Garreaud 2009). The prevailing westerly winds in Bolivia prevent moisture transport to the Andes during the austral winter (Vuille 1999). Three main sources of climate variability in Bolivia are the Pacific decadal oscillation, El Niño-Southern Oscillation, and Antarctic Oscillation (Seiler et al. 2013b).

Glaciers in Bolivia are situated in two mountain ranges in the outer tropics—Cordillera Occidental along the western

Fig. 1 **a** Altitudinal divisions in Bolivia. **b** Relative locations of the study sites (*A* Sajama, *B* Cololo and Huanacuni) with the Amazon basin



border with the Chile and the Cordilleras Apolobamba, Real, Tres Cruces and Nevado Santa Vera Cruz in the east (Vuille et al. 2008). Glaciers in the Cordillera Real such as Nevado Illimani were studied extensively by many researchers using ice core records and remote sensing (Ramirez et al. 2003; Ribeiro et al. 2013). Figure 1b shows the relative locations of the two study sites (A. Nevado Sajama, B. Nevado Cololo and Nevado Huanacuni) considered in this research with the Amazon Basin. Sajama ice-covered volcano (18° 06′ S, 68° 50′ W, 6542 m ASL) in the Oruro Department, Sajama Province in Bolivia is a stratovolcano in the Central Volcanic Zone (CVZ) in the Central Andes and is the highest peak in Bolivia and the southernmost ice-cap in the intertropical zone. Nevado Cololo (14° 50′ S, 69° 06′ W, 5859 m ASL) and the nearest Nevado Huanacuni (5058 m asl) in the Cordillera Apolobamba are separated from the Cordillera Occidental by the Altiplano is the other study site considered in this study. By using Landsat images, Oliveira (2013) observed that the Cololo-Huanacuni complex is currently not having any glaciers below 4626 m ASL whereas in 1975, the lowest glacier terminus observed was about 4317 m ASL.

Monthly mean precipitation rate is having high longitudinal gradient from west to east, and hence, the observed precipitation is very different in the two cordilleras in Bolivia. Precipitation occurs by the mechanism of Amazonian monsoon that causes about 80 % of the annual precipitation from October to April, particularly in the Cordillera Real (Rabatel et al. 2006). In the Sajama and Cololo regions, monthly precipitation pattern is similar but varies quantitatively (Fig. 2). It is also visible that the precipitation patterns are different during May–July. This means that there is a latitudinal gradient in precipitation from one study site to the other (Nevado Sajama is close to the Pacific as well as the subtropics). Higher precipitation rate occurred near the Nevado Cololo is due to the enhanced moisture content transported by the atmospheric circulation from the Amazon Basin. This process seems to be weakening towards the Nevado Sajama.

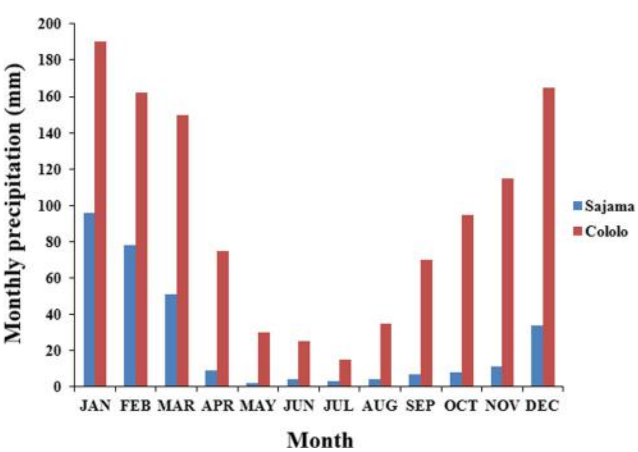


Fig. 2 Difference in the mean monthly precipitation (in mm) near the study locations in the two cordilleras in Bolivia

3 Datasets

Both remote sensing and meteorological datasets were used in this study. The remote sensing data includes multispectral images and digital elevation models from various sensors. The meteorological data includes temperature and precipitation data obtained from various sources. The second type also includes the three indices—ENSO, PDO, and AAO. The following subsections describe the datasets used and their suitability in this research.

3.1 Satellite data

A large number of research papers are available on the measurement of physical and surface characteristics of glaciers including snowline position and altitude, surface elevation, and terminus position based on remote sensing and photogrammetry (Arnaud et al. 2001; Bamber and Rivera 2007; Rabatel et al. 2012; Veettil et al. 2014). Satellite images are available since 1972 from Landsat series (MSS, TM, ETM+, and LDCM). Multisource image data including Landsat TM, ETM+, and IRS LISS III taken during the austral winter (May–August) were used in this research. During this time, ablation will be least and the glacier terminus would not be covered with fresh snow. Landsat TM and ETM+ images are having a spatial resolution of 30 m in visible and infrared channels. Spatial resolution of thermal channel in Landsat TM is 120 m and that in ETM+ is 60 m. A panchromatic channel (15 m) is also present in ETM+. LISS III is a multispectral camera operating in four spectral bands, three in the visible and near-infrared (VNIR) and one in the shortwave infrared (SWIR), each with a spatial resolution of 23.5 m. Spectral coverage of the images used here is given in Table 1. Only cloud-free images were used in order to avoid the difficulty in delineation of ice margin. Other than

Table 1 Spectral coverage of the images used

Channel	Spectral range (μm)		
	Landsat TM	Landsat ETM+	IRS LISS III
Blue (TM1)	0.450–0.520	0.450–0.515	
Green (TM2)	0.520–0.600	0.525–0.605	0.520–0.590
Red (TM3)	0.630–0.690	0.630–0.690	0.620–0.680
NIR (TM4)	0.760–0.900	0.775–0.900	0.770–0.860
SWIR1 (TM5)	1.550–1.750	1.550–1.750	1.550–1.700
Thermal IR (TM6)	10.40–12.50	10.40–12.50	
SWIR2 (TM7)	2.080–2.350	2.090–2.350	

multispectral images, DEMs from Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) Global Digital Elevation Models (GDEM) were also used in this research. The DEMs used were having a spatial resolution of 30 m and a vertical accuracy of 20 m which is suitable for the selected glaciers for the study due to comparable slopes. Co-registration of the images with DEMs and atmospheric correction using MODTRAN were done before applying other image processing algorithms. Image processing steps were done using Erdas Imagine and ESRI ArcGIS 10.1 software packages. Landsat TM image subsets (false color composites) of the study sites and the selected glaciers considered in this research are given in Fig. 3.

3.2 Meteorological data

High resolution, gridded, monthly precipitation and temperature (above 2 m from the ground level) data with a horizontal resolution of 0.5° lat-long during a period of 1948 to 2008 from the University of Delaware is used in this research (http://www.esrl.noaa.gov/psd/data/gridded/data.UDel_AirT_Precip.html). These data were derived from a large number of stations including Global Historical Climate Network (GHCN2) and archives of Legates and Willmott (Legates and Willmott 1990). Even though the precipitation and temperature data were derived from various existing meteorological stations, it is not totally free from errors. In mountainous environments, installation of meteorological stations is not always possible and the presence of altitudinal gradient in precipitation and surface temperature can also contribute some errors. Ocean Niño Index (ONI) and AAO index were downloaded from Climate Prediction Center (CPC), National Oceanic and Atmospheric Administration (NOAA) (http://www.cpc.ncep.noaa.gov). In this data, cold and warm episodes were defined when a threshold of ±0.5 °C is met for a minimum of five consecutive overlapping seasons. Pacific Decadal Oscillation (PDO), an index based on the variations in SST in the north Pacific, is downloaded from Joint Institute for the Study of the Atmosphere and Ocean (JISAO) (http://jisao.washington.edu/data_sets/pdo/).

4 Methodology and results

Image processing and meteorological data analysis were the two step processes done in this research. Image processing is used to calculate the snowline altitudes. Length (horizontal) changes in the terminus are not considered here because in the accumulation area it may or may not represent a change in the mass balance (Bamber and Rivera 2007). Many researchers studied the response of ELA to climatic variations in the Andes and Alps (Arnaud et al. 2001; Rabatel et al. 2012; Veettil et al. 2014). ELA separates the accumulation (positive mass balance) zone from the ablation (negative mass balance) zone and hence is a good proxy for monitoring mass balance changes. The highest SLA detected using satellite imagery during the dry austral winter can be considered as an equivalent to estimate the equilibrium line altitude of the year (Rabatel et al. 2012).

4.1 Calculation of snowline altitudes

Landsat TM3/TM5 and TM4/TM5 band ratio images (Bolch and Kamp, 2006) were widely used to distinguish "clean" glaciers from other land surface features by applying a suitable threshold to the ratio image, though ratio images are sensitive to the influence from thin clouds. Normalized Difference Snow Index (NDSI = [TM2 – TM5] / [TM2 + TM5]) is excellent for the spectral discrimination of snow and other surface features like soil, rock, dust, and water and is also proved to be suitable for snow cover mapping in rough topography (Silverio and Jaquet 2005). Even though band ratios and NDSI can be used to discriminate the spectral characteristics of snow and ice, it is still inadequate to calculate the snowline, particularly when the terminus is covered with fresh snow or a continuous ablation occurs throughout the year. Therefore, we did not adopt band ratios or NDSI methods in this research. Rabatel et al. (2012) adopted another methodology to calculate SLA using a Landsat 5-4-2 false-color composite image that can be used to get a more accurate value of the ELA and is followed in this study. In this method, TM4 (IRS 3) and TM2 (IRS 1) channels were applied with threshold values of 60 to 135 and 80 to 160 respectively before creating the 5-4-2 composite image to map the SLA successfully (4-3-1 for LISS III image). On interannual scales, when dealing with long-term climate influence, changes in the area of glaciers are not so useful compared to the equilibrium line or mass balance changes (Veettil et al. 2015) and hence not is used as a parameter in this study. While using remote sensing techniques, it is difficult to calculate error in the calculation of SLA and it depends on the image coregistration error in relation to the horizontal and vertical DEM resolution as well as depends on the terrain slope (Arnaud et al. 2001). Glaciers selected for calculating the snowline were free from shadow effects and were having lower slopes. However, the same method can be applied when shadow effects or steep terrain are present by using a suitable topographic correction.

The results of the SLA calculations at Nevado Sajama in the Cordillera Occidental and Nevado Cololo and Nevado Huanacuni in the Cordillera Apolobamba are summarized in Fig. 4. From the calculated values of SLAs during the study period, Nevado Sajama is found to have fluctuated heavily when cold and warm phases of ENSO occurs during the cold and warm regimes respectively of PDO (described in subsection 4.3). Apart from higher rates of precipitation, Nevado Cololo and Nevado Huanacuni also fluctuated in a

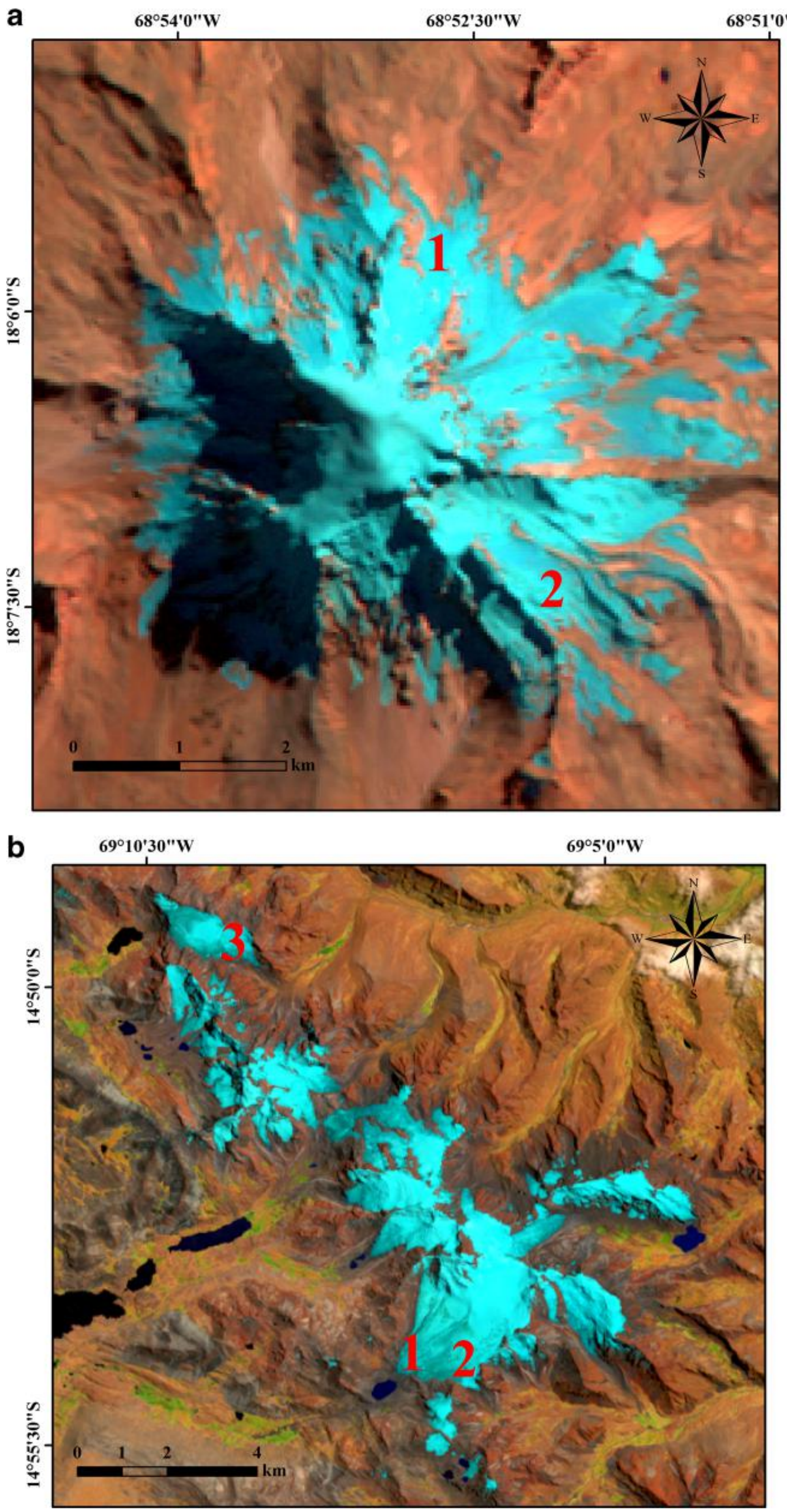


Fig. 3 Landsat false color composite image subsets of the study sites and their selected glacier outlets to calculate snowline altitudes of the study sites **a** Nevado Sajama and **b** Nevado Cololo (glacier outlets 1 and 2) and Nevado Huanacuni (outlet 3)

similar manner but the rate of recession in this region is higher. It is interesting to note that the response of SLAs with ENSO and PDO is higher in Bolivia, compared to that in Ecuador as calculated by Veettil et al. (2014). The possible reasons are as follows: (1) Bolivian glaciers are subjected to increased fluctuations in SLAs with SST variations compared to the

Ecuadorian glaciers or (2) errors in the calculation of SLAs in the case of Ecuadorian glaciers were interfered due to the absence of a unique precipitation season. These two possibilities are considered in the discussion section. In this research, we considered two geometrical parameters as well—glacier altitude and exposure. As observed from Fig. 4a, b, different glacier outlets of the same region showed similar retreating trend. However, there exists a small difference in the behavior of glacier outlet 3 (Nevado Huanacuni) in the second study site, and this can be due to the altitudinal differences. It is known that glaciers at lower altitudes are less resistant to mass loss (except sublimation) compared to those at higher

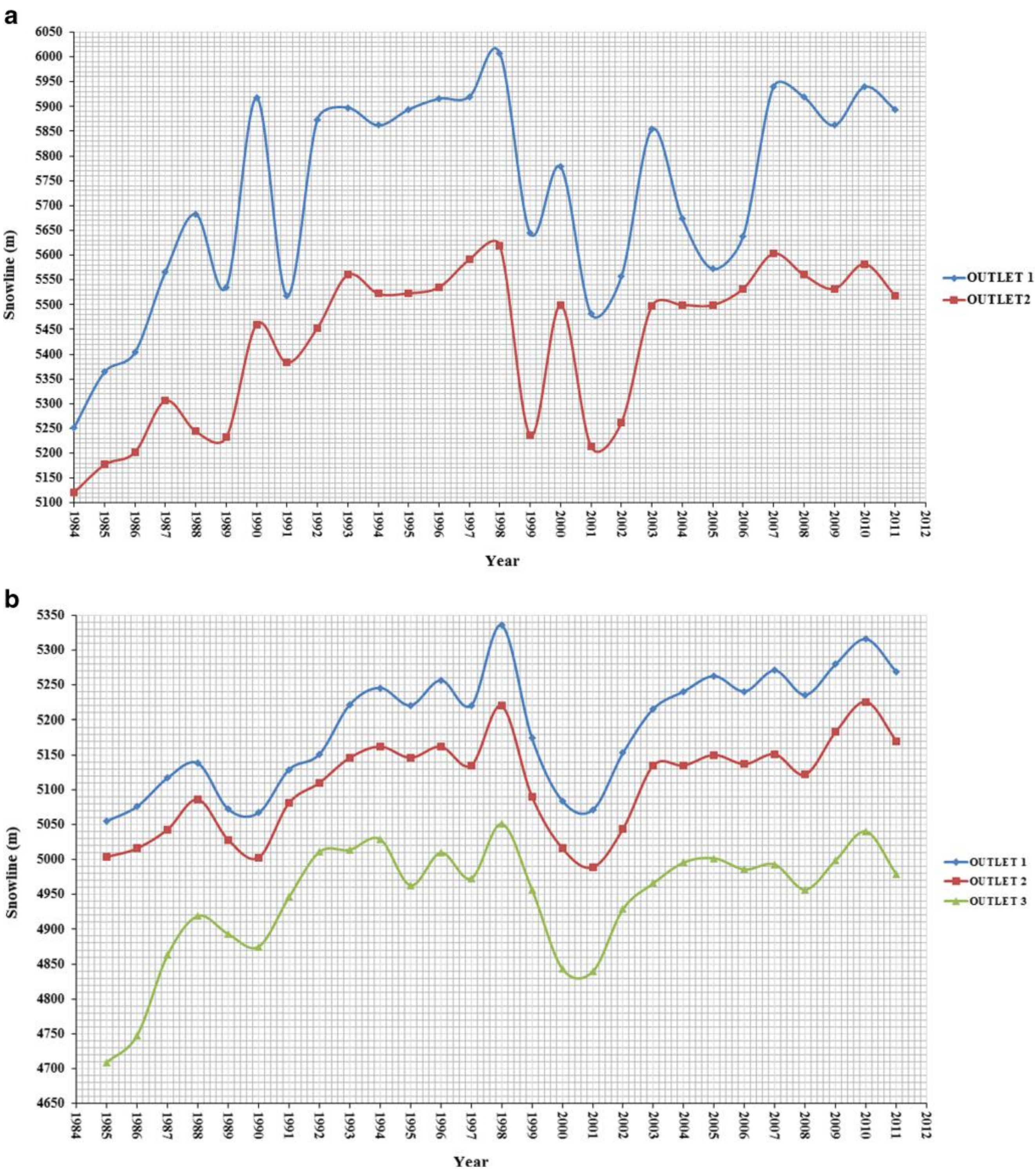


Fig. 4 Variations in the SLA (in meters) from 1984 to 2013. **a** Sajama and **b** Cololo and Huanacuni

altitudes, particularly in the lower and mid latitudes. This is visible in the case on glacier outlet 2 of Nevado Sajama during 2003–2007, when the glacier outlet 1 that is having a higher altitude has lowered its snowline during this period. Despite of altitudinal differences, snowlines of selected glaciers at same geographical locations varied in a nearly similar way.

4.2 Anomalies in precipitation and temperature in the study sites

We analyzed the anomalies in the monthly precipitation and temperature in the Cordillera Occidental and Cordillera Apolobamba in a similar way as described in Veettil et al. (2014) using the gridded datasets (1945–present) from the University of Delaware by applying linear interpolation in MATLAB. One cell each of the gridded datasets was used to calculate the anomalies for the two study sites. The weaker precipitation rate near the Nevado Sajama is clearly visible from Fig. 5a compared to that near the Nevado Cololo (Fig. 5b). The major precipitation source in the Central Andes is the moist air transported from the Amazon Basin. Accordingly, precipitation rate within the Amazon Basin is expected to be more in the eastern parts of the Altiplano than the western region and is visible from the Fig. 5a, b. During El Niño, moisture advection towards the Pacific is lowered, and this region is more sensitive to ENSO-induced circulation anomalies. Moreover, rapid fluctuations in the magnitudes of precipitation anomaly were observed at Sajama, and it is hypothesized that this geographic location is highly influenced by the changes in the Pacific SST. Note that this region is within 150 km from the Pacific coast. Unlike the anomalies in precipitation, temperature anomalies (Fig. 5c, d) are showing closely related patterns in the two study sites. It is seen that, in spite of the fact that the Sajama region is having a higher altitude, higher temperature anomalies were experienced during the strong El Niño period (during 1997–1998, for example). This shows that glaciers near the Pacific coast (in the western cordillera) might undergo fluctuations in ELA (and hence mass balance) with Pacific SST variations rapidly compared with glaciers in the Cordillera Apolobamba.

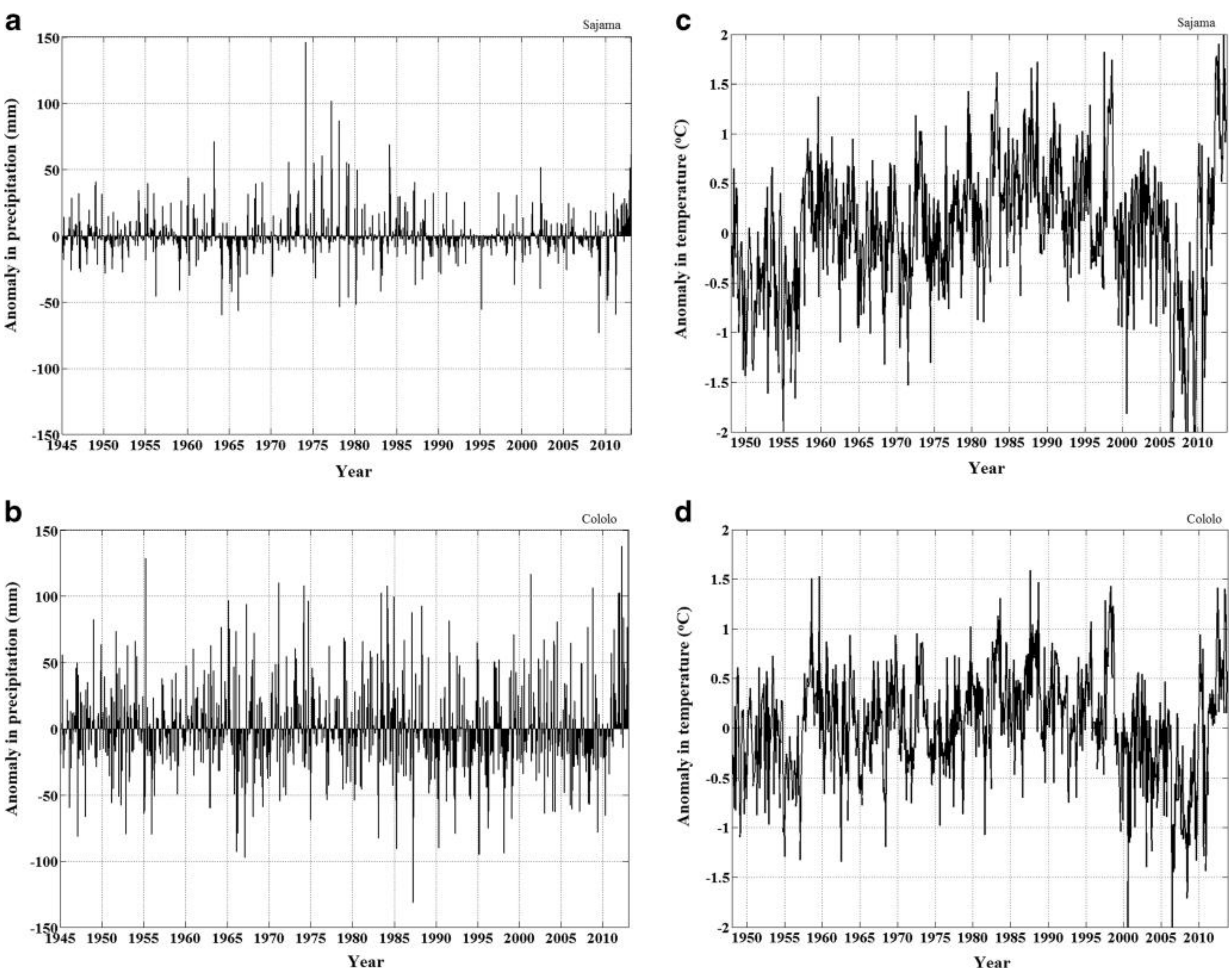


Fig. 5 Anomalies in precipitation (**a** Sajama and **b** Cololo) and temperature (**c** Sajama and **d** Cololo)

4.3 Considering the three sources of climate variability in Bolivia—ENSO, PDO, and AAO

We considered three sources of climate variability in Bolivia—ENSO, PDO, and AAO. A positive correlation between El Niño and PDO is observed, and El Niño-related precipitation anomalies were found to be stronger during the warm phases of PDO (Garreaud et al. 2009). Such a combination would influence the precipitation, if no changes in temperature occur throughout the year (which is common in the tropics), and snowline of the glaciers in this region might fluctuate accordingly. El Niño and La Niña events were known to affect the natural systems and economy in Bolivia for a long time. In this research, we considered the Antarctic Oscillation (AAO) as well, to investigate any connections existing between the positive and negative phases of AAO with the calculated changes in the snowline altitudes. AAO was observed to be causing tropospheric circulation variability towards the south of 20° S (Thompson and Wallace 2000). The geographical locations of the study sites, comparison to mountain glaciers in the inner tropics, are near to 20° S (Sajama, 18° S; Cololo, 14° S), and this justifies the inclusion of AAO in this study. The graphical representation of ENSO, PDO, and AAO indices are given in Fig. 6.

El Niño episodes are associated with below normal precipitation and warmer than normal conditions in the tropical region of the Andes. However, above normal precipitation is observed towards the southeastern region during El Niño conditions (Garreaud et al. 2009). The positive correlation between ENSO and temperature maximize during December–March (Garreaud et al. 2009) and the calculation of snowline during May–August is suitable because there is a lag of 1–3 months to occur an atmospheric response to ENSO in the continent (Kumar and Hoerling 2003). Moreover, the calculated snowline during this period gives a more accurate approximation of the ELA (Rabatel et al. 2012). The absence of

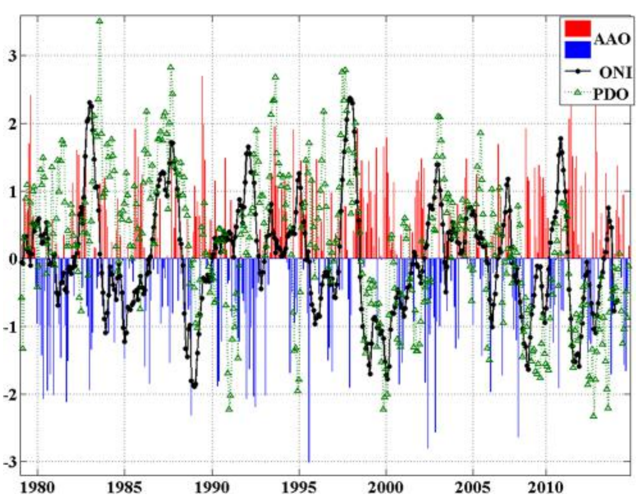


Fig. 6 ENSO, PDO, and AAO indices

seasonal snow and low ablation at the terminus during this period make it an easy task to calculate the snowline using satellite images. However, the influence of ENSO can be altered further due to other factors (strong westerly, for example) prevailing in the continent. Hence, it is difficult to establish a direct correlation between ENSO and PDO with the changes in snowline rather than observational investigation. Only a few station data are available, and this limits the calculation of interdecadal changes and spatial patterns of PDO, which is a long-lived pattern of pacific climate variability and shifts its regime in many decades. Even though there are signals of combined influence of ENSO and PDO on glaciers in the inner tropics (Veettil et al. 2014), the link between both stays unclear (Newman et al. 2003). Higher positive anomalies in precipitation are observed before 1980, before the PDO entered its recent positive regime until 2008. The influence of AAO is, however, visible significantly below 40° S (Gillett et al. 2006).

5 Discussion

Glaciers in different climatic conditions respond to similar climatic perturbations differently (Sagredo and Lowell 2012). In a warming climate, a high wintertime temperature can accelerate glacier ablation (Bonanno et al. 2013). Kaser and Osmaston (2002) proposed that if the location of a glacier is above the mean annual 0 °C isotherm, it would be highly sensitive to precipitation variability and insensitive to temperature variability. Summer accumulation glaciers are more sensitive to temperature variability than those with winter-accumulating ones (Fujita 2008). Larger surface mass balance variability found in the case of Zongo glacier in Bolivia shows the vulnerability of low-lying glaciers during the recent decades (Soruco et al. 2009). In a warming environment, when no change in precipitation occurs, smaller glaciers in the lower altitudes disappear faster (Chevallier et al. 2011) due to the lowering of accumulation/ablation ratio. It is reported that glaciers in Bolivia have been retreated rapidly between 1975 and 1983, and again between 1997 and 2006 (Soruco 2008), and this research confirms this finding. Periods of stability in the mass balance were observed during 1956–1975 (PDO cold regime) and 1992–1996 (unexpected). Similar trends were observed in the case of Antizana 15 (Francou et al. 2000, 2003) and Cotopaxi (Veettil et al. 2014) in the Ecuadorian Andes. In the case of Ecuador, the calculation of SLA which is close to ELA of the year is difficult to obtain due to the absence of a specific precipitation season, and this might cause additional errors in calculating the SLA along with the discrepancies in the remote sensing data. However, it is found that a rapid retreat has occurred in the case of Nevado Cololo, when compared with glaciers in Ecuador. Inner tropics are found to be getting cloudier and wetter than the outer tropics that are getting drier (Vuille et al. 2008). Snowline of Nevado Sajama fluctuated with the occurrence of ENSO (in phase with PDO) events, whereas the snowlines of Nevado Cololo and Nevado Huanacuni showed a (nearly) continuous increasing trend. The elevation in the snowline during the El Niño episodes during 1991–1995 must be evaluated carefully. There was a rapid increase in the snowline during 1991–1992 (Fig. 4a, b). This snowline did not vary much during 1992–1995 even though the El Niño events were stronger and prolonged. One of the possible reasons behind this anomaly can be the cooling effect of the volcanic aerosols in the stratosphere due to the eruption of Mount Pinatubo on June 15, 1991 (Rabatel et al. 2013).

Glaciers in the tropics are having two special features—these are subjected to higher levels of energy forcing due to the specific latitudinal and altitudinal location; and accumulation and ablation occurs simultaneously (this equilibrium can be broken by climate imbalances) due to year-round precipitation (Chevallier et al. 2011; Kaser and Osmaston 2002). The influence of ENSO and other ocean-atmospheric phenomena on the Andean climate varies along its length (Garreaud 2009). In the outer tropics, the annual distribution of precipitation, particularly during December–February, influences the annual melting (Favier et al. 2004). However, the precipitation variations take a longer time to affect the glacier terminus in comparison with temperature fluctuations (Bonanno et al. 2013). Many researchers (Raper and Braithwaite 2006; Rupper and Roe 2008) suggest that the glaciers in wet climate are more sensitive to temperature rise than in a dry climate. Main moisture source for precipitation in the Altiplano is transported from the eastern lowlands of the Andes, which is highly dependent on the tropical SST anomalies (Vuille et al. 2000). El Niño events induce precipitation deficits in the outer tropics, which in turn promotes glacier melting. In the inner tropics, temperature increase associated with warm phases of ENSO events causes augmented melt rates. Moreover, rate of liquid precipitation is higher in the inner tropics whereas solid precipitation is almost absent, which is common in the outer tropics (Favier et al. 2004). Other than altitudinal variations, the rate of glacier mass loss in the tropical Andes is highly dependent on the geometric characteristics such as exposure. Glaciers oriented towards east and south were found to have lesser retreat compared with those oriented towards the north and west (Soruco et al. 2009). The slight mass gained by these glaciers during 1963–1975 can be explained by the predominant cold regime of PDO during 1962–1976 (Veettil et al. 2014). It is seen that there is an imbalance in the frequency of occurrences of PDO after 1998–1999. Warm phases of PDO are observed to be dominating recently until 2007.

ENSO phase changes can have an important role in modulating the AAO phases (Carvalho et al. 2005). Even though most of the studies pointed out that AAO has its significant effects on the extratropical climate (Carvalho et al. 2005; Pohl et al. 2010), a recent study (Gong et al. 2009) has pointed out that there are signatures of AO and AAO in the tropical coral proxies over the South China Sea. Antarctic climate changes are neither totally independent on Pacific SST variations nor a regional issue (Carvalho et al. 2005). AAO is one of the causes of atmospheric variability in the southern hemisphere up to higher latitudes (20° S) (Pohl et al. 2010), and ENSO affects stratospheric circulation variations, both in the tropics and poles (Hurwitz et al. 2011). It is seen (Carvalho et al. 2005) that majority of the positive AAO conditions were associated with a negative ENSO (La Niña) events and vice-versa, which shows that there are some interrelations between ENSO and AAO. But other than ENSO and its combined mode with PDO, no significant influence was observed on the snowline variations due to the phase changes of AAO. However, further studies are needed for clarification. It might be interesting to investigate whether the occurrence of a positive phase of AAO is to recover the combined warming effects of ENSO and PDO.

Inter-annual variability of the mass balance at the study sites are highly dependent on the inter-annual variability of the precipitation during the summer (Favier et al. 2004) and the interannual precipitation variability is highly dependent on the atmospheric circulation anomalies during the extreme phases of Southern Oscillation (Vuille et al. 2000). Sometimes, a snow-cover of higher albedo is enough to prevent the rapid ablation during the time of weak precipitation (Soruco et al. 2009). Precipitation and air temperature are the determining factors in the SLA changes in the outer tropics. Precipitation and temperature at the study sites were highly influenced by ENSO due to the reduction in the advection of moist air from the continent due to powerful mid-level westerlies (Vuille 1999; Vuille et al. 2000; Garreaud and Aceituno 2001). Due to low spatial resolution of the precipitation datasets, correlation between ENSO and precipitation variability was not established in this research. Moreover, installation of meteorological stations at very high altitudes is nearly impossible, and this would add some errors to the calculated anomalies in precipitation and temperature from gridded datasets with a resolution of 0.5 × 0.5°. Precipitation near the Nevado Sajama is highly influenced by the large-scale circulation changes over the Altiplano (Vuille 1999). Correlation between the ENSO and the precipitation in the Altiplano is reported to be highest during December-January-February (Garreaud et al. 2009). During the occurrences of the El Niño, westerly wind anomalies prevent the transportation of moisture from the eastern continent into the Sajama region (Vuille 1999). In the case of Nevado Cololo, a higher monthly precipitation occurs (Fig. 2) due to the moist air transport from the Amazon Basin. It is known that dry and cold conditions prevail along the pacific coast, and relatively warm and humid conditions prevail in the continent (Garreaud 2009). Due to this dry and cold condition near the Nevado

Sajama, variation in its SLA is not as rapid as the case of Nevado Cololo. In addition, western side of the Altiplano is more sensitive to ENSO than the eastern region (Vuille et al. 2000). Walker circulation is one of the defining features of tropical climate—a stronger Walker circulation implies a La Niña condition and a weaker one implies an El Niño condition. Ice core records from Quelccaya ice cap in Peru suggest that a rise in temperature has occurred during the twentieth century (Thompson et al. 1984), even though the changes in precipitation are not so visible (Vuille et al. 2008). Even though there is no relevant documentation on the humidity records exist in the Andes, Vuille et al. (2008) could found a moderate rise in relative humidity between 1950 and 1995 in the Western Bolivia based on station data. One of the drawbacks in the study of ENSO impact on tropical glaciers in the past is that an advance (or a retreat) cannot be verified always due to the uncertainties in moraine dates and lack of satellite imagery or aerial photography.

6 Conclusions

Even though unexpected snowfall events sometimes cause the underestimation of ELA from SLA using satellite imagery, the selected images towards the end of dry season were excellent to calculate the snowline. In the outer tropics, the highest snowline during the dry season can be taken as the ELA of the year. From the results from this study, it is seen that the snowline of the Nevado Sajama, Nevado Cololo, and Nevado Huanacuni have been fluctuated between warm and cold phases of ENSO (combined with the warm and cold regimes of PDO, respectively) between 1984 and 2011. The significant snowline rise of the selected glaciers during 1990–1995 confirms the effect of El Niño, which persisted for a long period in phase with the warm PDO. In general, there was an overall increase in the snowline during this period, and this indicates that the climate condition in Bolivia is still warming. For a better understanding of this complex glacio-climatological process, other local climatic factors such as seasonal snowfall, wind speed, and sublimation characteristics of the glacier can be considered for future improvement.

Acknowledgments First author likes to acknowledge Rio Grande do Sul State Foundation for Research (FAPERGS) for providing his PhD research scholarship and Ms. Geana Veiga Aurelio for the help with statistical data analysis. We would like to thank USGS, NOAA, and JISAO for the datasets used in this research.

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CAPITULO VII

Variations in annual snowline and area of an ice-covered stratovolcano in the Cordillera Ampato, Peru, using remote sensing data (1986-2014)

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Geocarto International,31, 544-556, 2016

Geocarto International, 2016
Vol. 31, No. 5, 544–556, http://dx.doi.org/10.1080/10106049.2015.1059902

Variations in annual snowline and area of an ice-covered stratovolcano in the Cordillera Ampato, Peru, using remote sensing data (1986–2014)

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(*Received 18 April 2015; accepted 26 May 2015*)

This research focuses on the recent variations in the annual snowline and the total glaciated area of the Nevado Coropuna in the Cordillera Ampato, Peru. Maximum snowline altitude towards the end of dry season is taken as a representative of the equilibrium line altitude of the year, which is an indirect measurement of the annual mass balance. We used Landsat and IRS LISS3 images during the last 30 years due to its better temporal coverage of the study site. It is found that there was a decrease of 26.92% of the glaciated area during 1986–2014. We calculated the anomalies in precipitation and temperature in this region and also tried to correlate the changes in glacier parameters with the combined influence of El Niño – Southern Oscillation (ENSO) and pacific decadal oscillation (PDO). It is concluded that the snowline of Nevado Coropuna has been fluctuated during ENSO, and maximum fluctuations were observed when ENSO and PDO were in phase.

Keywords: snowline; outer tropics; equilibrium line altitude; Nevado Coropuna; Cordillera Ampato

1. Introduction

Tropical glaciers in South America are observed to be shrinking since the Little Ice Age. A significant change in the tropical Andean climate is observed during the past few decades (Vuille et al. 2008). It is predicted that all tropical Andean glaciers at lower latitudes will disappear soon (Veettil et al. 2014), and this glacier recession is dependent on the climate variations. Changes in climate induces changes in atmospheric humidity, precipitation and cloudiness towards which the glaciers are sensitive (Vuille et al. 2008). The Andes is considered as the most important mountain range in the southern hemisphere like the Alps or the Himalayas in the northern hemisphere. It is observed that the percentage of warmer nights have been increased since the second half of twentieth century in South America, and this warming trend is not homogeneous in the eastern and western slopes of the Andean chain (Salzmann et al. 2009). Many people in the hyper-arid coastal lowlands and Altiplano in Bolivia and Peru rely on glacier meltwater for their freshwater needs during the dry season and

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hence mountain glaciers can be considered as freshwater buffers in this region. Tropical mountain glaciers are considered as indicators of climate change as they grow or shrink with climate imbalances (Vuille et al. 2008). Tropical Andes (between 10°N and 16°S) is one of the suitable locations to study the climate change influence on tropical glaciers because these glaciers are subjected to higher daily temperature variations than annual temperature variation. Vuille et al. (2000) proposed that the rainfall variability towards the east of the equatorial Andes related more to the tropical Atlantic circulation anomalies than those in the Pacific. Another point on precipitation variability in the Andes is that the precipitation rate increases with altitude and observed particularly in the high altitude Ecuadorian Andes (Garreaud 2009). Precipitation is one of the variables that determine the growth/mass loss of glaciers in the tropical region and it is worthwhile to understand how the precipitation varies in this region of the Andes. It is also noted that the precipitation near the glaciers within the Amazon basin is higher compared to those near the Pacific (Veettil et al. 2014). The anomalies in temperature and precipitation associated with El Niño – Southern Oscillation (ENSO) were found to be weakening towards the Altiplano (Garreaud 2009).

The Peruvian Andes accommodates about 70% of the tropical glaciers (Vuille et al. 2008) and is the most extensive ice-covered tropical mountain range. An increase of about 0.1 °C per decade in the air temperature is reported in the central Andes (Vuille et al. 2008). The land cover in Peru can be divided into three geographical regions: Pacific coast, Cordillera of the Andes and the Amazonian forest (Chevallier et al. 2011). The Peruvian climate is highly influenced by the Andean mountain chain. Changes in the equilibrium line altitude (ELA) denote an immediate change in the mass balance of the glacier and a continuous change in ELA can be used to estimate the climate trend in that region on an interannual scale. It is identified that the ENSO and other global scale phenomena such as the pacific decadal oscillation (PDO) influence the Andean climate differently along its length (Garreaud 2009; Veettil et al. 2014). ENSO dominates in the Tropical Pacific in the southern hemisphere whereas PDO dominates in the North Pacific. It is known for ages that El Niño years were followed by decreased precipitation over northern South America. This is due to the inhibition of moisture transport from the Amazon basin due to strong westerly wind during El Niño (Vuille 2013). It is also noted that the river discharge during strong El Niño periods were higher towards northern Peru compared to those in the south (Lavado-Casimiro et al. 2013). Unfortunately, many of the hydrometeorological stations in Peru are discontinuous or not existing at present. Ice core records from Peru show a direct correlation between ENSO and glacier mass balance in the Peruvian Andes (Henderson et al. 1999; Thompson 2000; Herreros et al. 2009). Many papers are available on the fact that smaller glaciers in the Cordillera Blanca in Peru are disappearing recently (Racoviteanu et al. 2008; Salzmann et al. 2009; Rabatel et al. 2013). The Quelccaya ice cap in Peru is having one of the well-documented ice core records in the world (Thompson 2000).

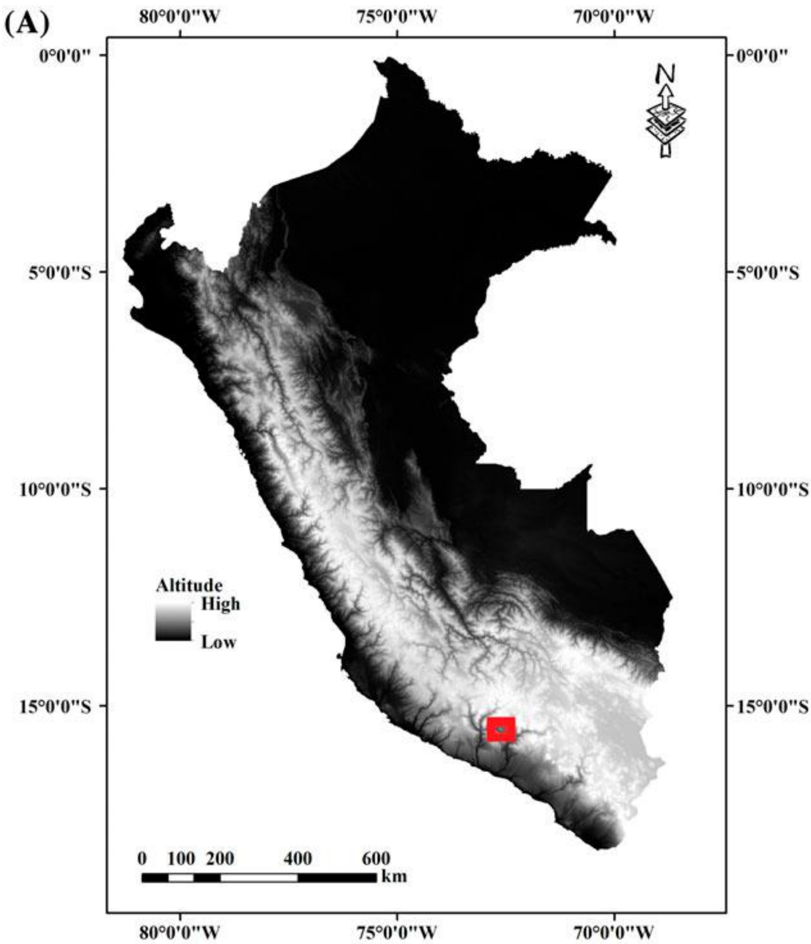
Due to the availability of SPOT and Landsat series of images, it is possible to understand the overall decline in the glaciarized area in the Cordillera Blanca since the early 1970s (Vuille et al. 2008). Studies based on remote sensing have predicted that many glaciers in Peru, such as Yanamarey Glacier, would disappear within a decade (Huh et al. 2012). In this study, we calculated the annual changes in the area and annual snowline of a glaciated stratovolcano – Nevado Coropuna – in the Cordillera Ampato of Peru and tried to understand how these two glacier parameters changed with the phase changes of ENSO and PDO.

2. Study site and climate conditions

The Nevado Coropuna (Lat: 15°24′-15°51′ S; Long: 71°51′-73°00′ W) in the central volcanic zone (CVZ) in the Cordillera Ampato, southern Peru, is considered in this research (Figure 1). Cordillera Ampato is consisted of 93 glaciers with an average thickness of about 35 m and a total surface area of 146.73 km^2 based on aerial photography in 1962. The Nevado Coropuna is the highest peak (6426 m asl) in the Cordillera Ampato and the highest stratovolcano in Peru (Racoviteanu et al. 2007). Many people in the northern–western part of Arequipa city depend on the glacier melt water supply from Nevado Coropuna. Recent glacier shrinkage in the Cordilleras of Peruvian Andes is reported to be started in the second half of 1980s (Salzmann et al. 2012). Racoviteanu et al. (2007) found that the size of the Coropuna was decreasing from 82.6 km^2 in 1962 to 60.8 km^2 in 2000. Mass balance of glaciers in this region depends highly on the variations in precipitation (Wagnon et al. 1999).

Precipitation in the Coropuna region depends mainly on the easterly circulation of air masses from the tropical Atlantic Ocean (Herreros et al. 2009). However, the Pacific atmospheric circulation patterns are also having a significant role in determining the climate in this region. Like other glaciers in the subtropics and outer tropics, Nevado Coropuna is also having an ELA above the 0 °C isotherm, whereas those in the inner tropics are close to the 0 °C isotherm. Glaciers situated in the outer tropics and the subtropics are thus considered as temperature-insensitive (Kaser 1999). The east of the Atlantic Ocean and the Amazon basin control the precipitation in the tropical Andes, mainly by the seasonal easterly winds (Vuille & Keimig 2004). There are 15 meteorological stations operated by the Peruvian national meteorological and hydrological service within 60 km of the study site. Seasonal variation in temperature is small whereas that in precipitation is higher and about 70–90% of the precipitation occurs during the austral summer (December–March). Dry season in the tropical Andes of Peru is during the austral winter. Higher precipitation rates were observed on the east-facing slopes than the west-oriented ones, probably due to higher moisture transport from the Amazon basin. Decreased precipitation rates were observed during strong El Niño events during 1982–1983 and 1992 whereas a strong El Niño in 1997 was found to be not interfered with the observed precipitation rates (Herreros et al. 2009). Unfortunately, many of the meteorological and hydrological stations in this region stopped functioning or having incomplete data-sets. Figure 2 shows the monthly mean precipitation (MMP) at the Coropuna region derived from various meteorological stations near the Nevado Coropuna.

Various studies focused on monitoring the changes in glaciated area, ice volume or ice thickness of the Nevado Coropuna using remote sensing and GIS techniques (Racoviteanu et al. 2007; Peduzzi et al. 2010; Ubeda 2011). There are previous studies on the snowline variations of the Nevado Coropuna and other glaciers in the central Andes during the late Pleistocene (Bromley et al. 2009, 2011). The effects of recent warming in the tropics on the Nevado Coropuna using ice core records was published recently by Herreros et al. (2009). In this study, we used remotely sensed satellite images and digital elevation models (DEM) for monitoring the changes in the snowline altitude (SLA) of the Nevado Coropuna and also to calculate the annual changes in the area towards the end of the dry season (May–September). We also used precipitation and air temperature data from the University of Delaware.



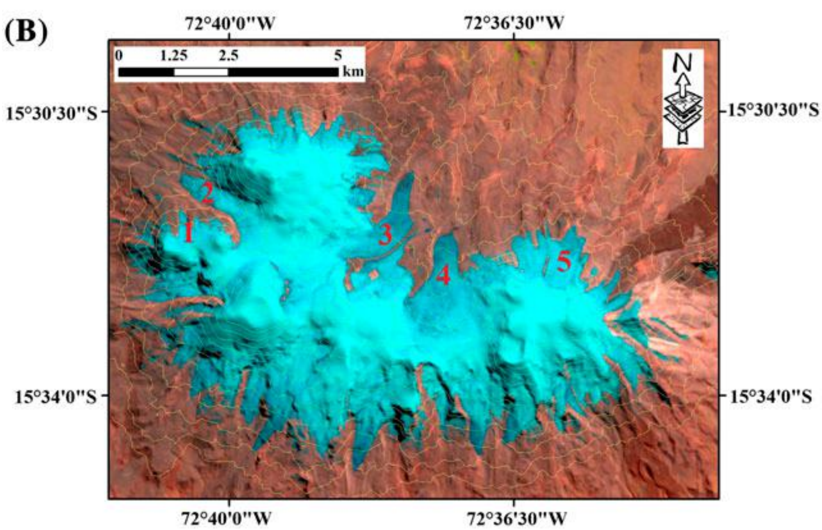


Figure 1. (A) Altitude distribution in Peru and the location of the Nevado Coropuna. (B) Selected glaciers for calculating the highest annual snowline.

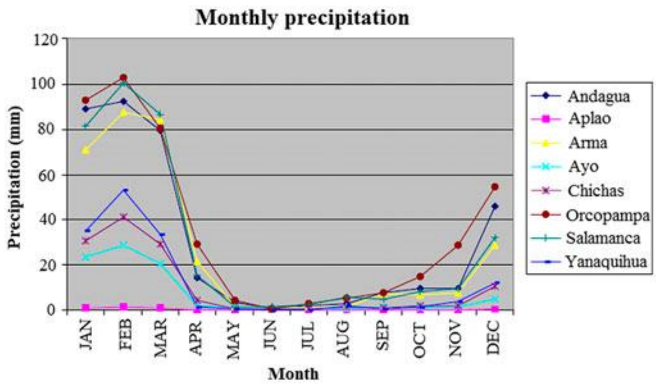


Figure 2. MMP measured by the meteorological stations near the Nevado Coropuna.
Source: Silverio 2005.

3. Data-sets

Landsat series of data are proved to be excellent in monitoring glacier surfaces, and many studies exist on the application of Landsat data in glaciology (Aniya et al. 2000; Albert 2002; Bamber & Rivera 2007; Racoviteanu et al. 2008) . Landsat bands in the VNIR contain the most valuable information to distinguish glacier surface characteristics (Pope & Rees 2014). Landsat series images (except Landsat 1–4) are having a spatial resolution of 30 m in the multispectral channels and only these wavelengths are used in this research. We also used an LISS3 image (spatial resolution: 23.5 m) from IRS-P6 in 2012 due to the discontinuity of the Landsat data in the same year. Other than multispectral images, we used DEM from ASTER GDEM V2 for calculating the surface properties of the study site. The DEM used here is having a vertical accuracy of approximately 20 m, which is having the same order of magnitude as the surface lowering and hence an altitudinal correction was not necessary (Rabatel et al. 2012). Moreover, the artefacts in SRTM caused by the penetration of C band in to the snow is absent in the ASTER GDEM.

In order to understand the influence of variabilities in the precipitation and temperature on the glacier mass loss in the study site, we used precipitation and temperature data from multiple sources for comparison. We used meteorological data from the University of Delaware in the form of gridded data-sets as well as from Servicio Nacional de Meteorología y Hydrología del Perú (SENAMHI). The gridded data-sets from Delaware are having a lat-long resolution of $0.5° \times 0.5°$. Ocean Niño Indices and PDO indices were downloaded from NOAA (http://www.cpc.ncep.noaa.gov).

4. Methodology and results

Identification of snow and ice using remote sensing is easy in theory, but in practice, it is not so straightforward (Albert 2002). This is because the spectral response of snow and ice varies with the quantity of impurities and meltwater above the ice and albedo changes with the ageing. Two satellite systems used for the last few decades, Landsat and SPOT, suffered from spatial and spectral resolution constraints, respectively. One of the drawbacks of using satellite images is that specific algorithms may be needed for images from place to place or season to season. Methods using Landsat series of images vary from simple false-colour composite and band ratios to principal component analysis, indices and object-oriented image analysis. In order to assure the accuracy and reduce the error, atmospherically corrected images were co-registered before further processing. One of the most efficient methods to map glaciers is the manual delineation. However, manual delineation is not applied here because a large number of images were used. We calculated the annual changes in the minimum area during the dry season and the maximum snowline from 1986 to 2014 to understand the influence of climatic perturbations on the Nevado Coropuna. In order to understand the glacier recession due to climate forcing, some of the variables such as atmospheric circulation and anomalies in temperature and precipitation are relevant. We calculated the anomalies in temperature and precipitation at the study site during the last 50 years and also considered two ocean-atmospheric phenomena in the Pacific – ENSO and PDO.

4.1. Variations in the annual minimum area of the Nevado Coropuna

The area of the glaciated surface and SLA can be used to understand the influence of climate change on glaciers in the outer tropics and subtropics. We tried to make sure that the images used were acquired towards the end of dry season (May–September) and images are cloud-free. In order to discriminate ice and other objects, we calculated normalized difference snow indices (NDSI) from Landsat and LISS3 images (Equations 1 and 2, respectively). NDSI images were calculated from the green (TM2: 0.52– 0.60 μm) and mid-infrared (TM5: 1.55 – 1.75 μm) channels.

$$\mathrm{NDSI} = [(\mathrm{TM2} - \mathrm{TM5})/(\mathrm{TM2} + (\mathrm{TM5})] \tag{1}$$

$$\mathrm{NDSI} = [(\mathrm{Band1} - \mathrm{Band4})/(\mathrm{Band1} + \mathrm{Band4})] \tag{2}$$

Glacier area can be calculated by applying a suitable threshold to the NDSI images. The threshold value to be applied to delineate glacier margin may vary from place to place and even from image to image (Wang & Li 2003). In this research, we used a threshold between 0.45 and 0.55 for Landsat images and 0.75 to 0.85 for the LISS3 images. By applying a suitable threshold, NDSI images can be used even when thin clouds are present in the image (Sidjak & Wheate 1999). Due to its robustness and easiness to apply, particularly when a large number of images are to be processed, Racoviteanu et al. (2008) used the NDSI method in the Cordillera Blanca. The observed changes in the annual minimum area of Nevado Coropuna during 1986–2014 are graphically represented in Figure 3.

It is seen from the Figure 3 that there was a decrease of 26.92% of the total glaciated area of the Nevado Coropuna from 1986 to 2014. There was a rapid decrease in the area during the El Niño episodes during 1997–1998, 2004–2005 and 2009–2010. It is also noticed that there was an increase in the area during the La Niña episodes in 1998–1999, 2001–2002 and 2010–2011. This indicates that the response of mountain glaciers to the ENSO in the Cordillera Ampato is rapid (whereas glaciers in Bolivia or Ecuador were found to show a delayed response towards ENSO). However, length (horizontal) changes in the terminus may or may not represent a change in the mass balance when the ice thickness is unknown (Bamber & Rivera 2007) or the case of surging glaciers (Aniya et al. 2000) and hence we considered another parameter – annual maximum snowline – which is given in the next subsection. From a glaciological viewpoint, aerial changes are more useful on decadal scales than annual basis when considering a long-term global climate change influence.

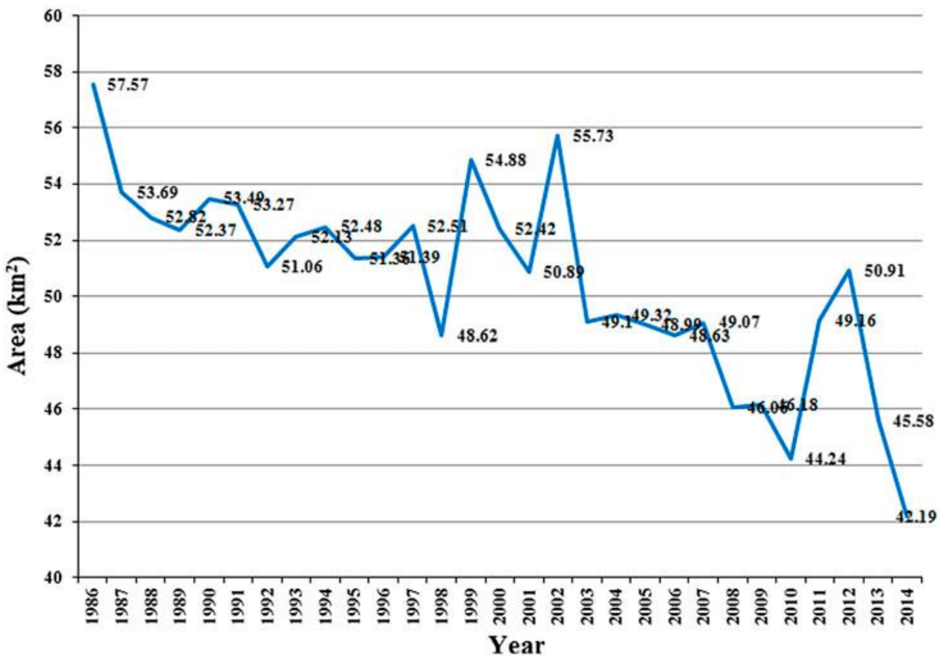


Figure 3. Changes in the annual minimum area of the Nevado Coropuna (1986–2014).

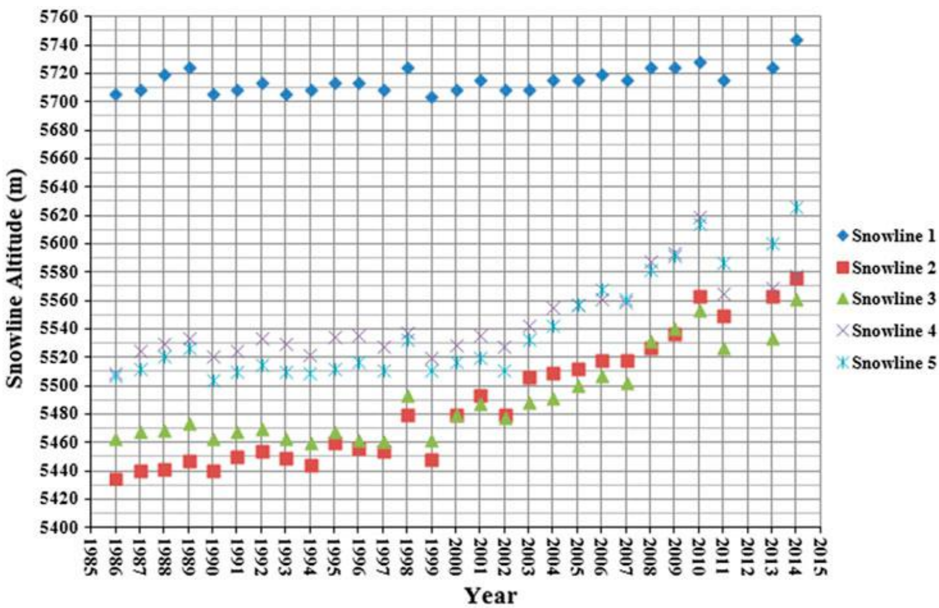


Figure 4. Variations in the annual snowline of selected glaciers (1986–2014).

4.2. Variations in the annual snowline maximum during dry season

The SLA was calculated based on Rabatel et al. (2012). The highest SLA calculated towards the end of dry season (May–September) can be taken as a representative of the ELA (Rabatel et al. 2012), particularly in the outer tropics and the subtropics. Based on comparison and validation with field data on Glaciar Zongo in Bolivia and Glaciar

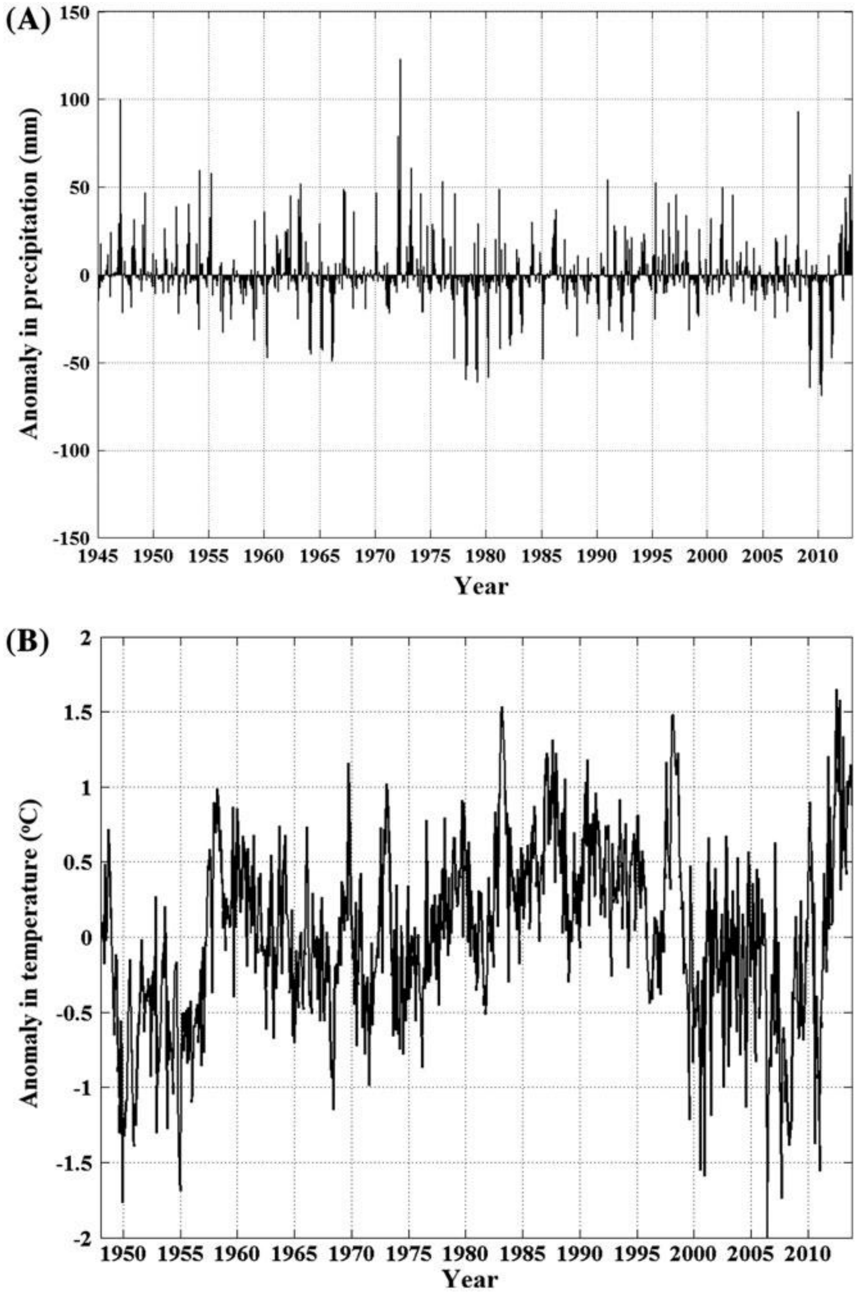


Figure 5. (A) Anomaly in precipitation; (B) Anomaly in temperature.

Artesonraju in Peru, it is found that SLA is a good proxy for ELA and hence can be used to measure annual mass balance changes (Rabatel et al. 2012). In contrast to inner tropics, there is a strong seasonality of precipitation in the outer tropics and hence we can use the maximum SLA during the dry season as a representative of ELA of the year. In order to calculate SLA of the selected glaciers (Figure 1(B)) from Landsat images, 5-4-2 false-colour composite images were created and certain threshold values were applied to TM2 and TM4. The threshold applied to TM4 may vary from 60 to 135 and for TM2, it vary from 80 to 160. The calculated annual snowlines of selected glaciers of the Nevado Coropuna during 1986–2014 are given in Figure 4. It is seen that there was an increasing trend in the SLA during this period and the snowline fluctuations were 'disturbed' with the ENSO and PDO phase changes. It is also seen that the glacier with the highest SLA studied (snowline 1 > 5700 m) showed less fluctuations compared to others due to high altitude, which is normal at higher altitudes due to lower temperature and higher snowfall. The highest snowline may not be at the end of the hydrological year (Rabatel et al. 2012) and it depends on seasonal snowfall, if occurred.

4.3. Calculating the anomalies in precipitation and temperature at the study area

We calculated the anomalies in the precipitation and the temperature near the Neva proven do Coropuna since 1950s using the data-sets from the University of Delaware (Figures 5(A) and (B)). We applied linear interpolation to the gridded data-sets in MATLAB to plot the anomalies. We used only one cell each for temperature and precipitation, which covers the entire study site, to plot the anomaly due to its resolution (0.5° × 0.5° lat-long). We also plotted the ENSO and PDO indices (Figure 6) during 1979–2014 to observe whether the changes in precipitation and temperature at the study site varied with the phase changes of ENSO and PDO. It is noted that a positive regime of PDO has prevailed from the late 1970 to 2008 then started an interrupted cold regime, which is not usual compared to normal patterns of PDO that persists for decades.

A strong correlation between ENSO (and PDO) indices and the calculated anomalies in precipitation is absent in the study region. However, the changes in the annual

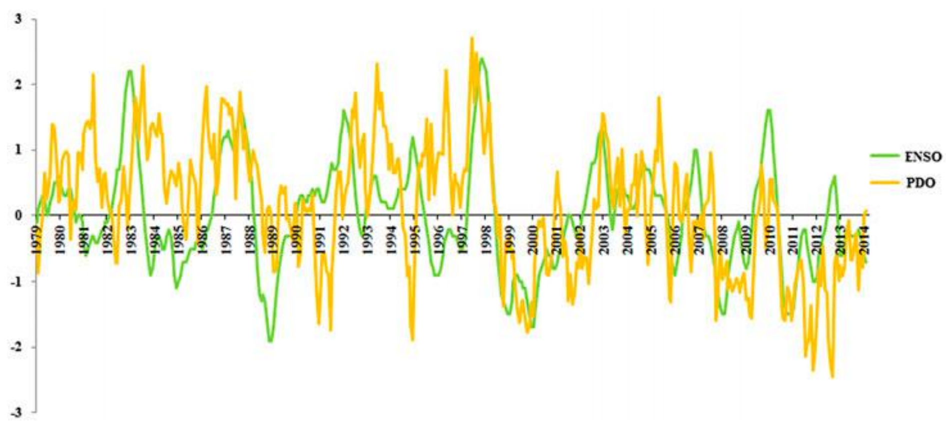


Figure 6. PDO and ENSO indices (1979–2014).

snowline and area of the glaciated region fluctuated heavily with the phase changes of ENSO and PDO and the temperature anomaly has also shown variations during strong ENSO conditions.

5. Discussion

The overall decrease in the glaciated area of the Nevado Coropuna during 1986–2014 (26.92%) is strikingly similar to that during 1962 to 2000 (26%) calculated by Racoviteanu et al. (2007) and this shows that the glaciated area of Nevado Coropuna was decreasing at the rate of about 26% during the last 50 years. Significant glacier recession in the Peruvian Andes started in the mid-nineteenth century itself (Kaser 1999). Some glaciers in the Cordillera Vilcanota in Peru were reported to have lost about 32% of the glacier area during 1962–2006 (Salzmann et al. 2012) and a loss of 35% in the southern part of Cordillera Blanca between 1962 and 1999 is also reported (Mark & Seltzer 2005). Kaser et al. (1996) studied the relationship between the changes in the ELA of Andean glaciers and fluctuations in the climate. However, the presence of excess snow cover on the glacier terminus was always a hindrance for delineating the glacier boundary and ELA using remote sensing data in this region and this problem was later overcome by using ‘snowline’ as a climatic indicator (Arnaud et al. 2001) for this type of glaciers. The presence of distinct dry and wet seasons makes it possible to calculate annual snowline more accurately using satellite images in the outer tropics compared to the inner tropics. Satellite images taken during the end of dry season (May–September) have already proven to be excellent in calculating the annual snowline which in turn can be used as substitute for the ELA based on the case studies on Glaciar Zongo, Bolivia and Glaciar Artesonraju, Peru, by Rabatel et al. (2012). The exceptional increase in the annual snowline and decrease in the surface area of the glacier during 1997–1998, 2004–2005 and 2010 can be well explained by the presence of strong El Niño occurred during the warm regime of PDO. From the results, it is seen that both the glacier area and the snowline did not vary much as expected from the combined influence of the El Niño and the positive PDO occurred during 1991–1995. The possible explanation for this zero or slightly positive mass balance can be explained on the basis of Rabatel et al. (2013) in such a way that the cooling effect of the volcanic sulphate aerosols in the stratosphere due to the eruption of Pinatubo interrupted the influence of the long El Niño during 1991–1995. However, this explanation is difficult to prove statistically.

The influence of ENSO on the ice-covered Nevado Sajama in Bolivia was reported in 2001 (Arnaud et al. 2001), which is located in the CVC where the Nevado Coropuna is also situated. Based on general circulation models (GCM), Minvielle and Garreaud (2011) calculated a significant decrease in the easterly circulation over the Altiplano that may cause a strong decrease in the precipitation in the tropical central Andes towards the end of twenty-first century. Inter-decadal variability in the Andes is associated with long-term changes in the pacific circulation patterns whereas decadal variability is associated with the changes in the circulation patterns over the Amazon basin (Espinoza Villar et al. 2009). Even though many models exist, such as CMIP3, the inconsistent trend in regional precipitation severely limits the understanding of climate changes over the Altiplano (Minvielle & Garreaud 2011). However, it is already understood that the Pacific SST has been increased since the late 1970s because of the so-called Pacific climate shift and this might be one of the causative agents of the observed accelerated glacier retreat (Rabatel et al. 2013). Glaciers situated in different climatic regimes can respond to similar climatic perturbations with different magnitudes (Sagredo & Lowell 2012). In the tropical Andes, this difference in magnitude of response is very visible because the climate is influenced by Atlantic, Pacific and Westerly circulation in varying magnitudes. This magnitude of the influence of the circulation patterns vary from the inner to the outer tropics and results in different MMP and temperature patterns. Altitude is also another important factor that controls the glacier recession in response to climate change (Chevallier et al. 2011). The rate of increase/decrease in the snowline situated at higher altitude (snowline 1) was found to fluctuate less compared to those at a lower altitude (snowline 2 and snowline 3). This is because the higher altitudes in this region are fed by heavy snowfall (note that the maximum precipitation occurs in the summer in low tropical latitudes) whereas rapid ice melting occurs in the lowest parts (Chevallier et al. 2011). At higher altitudes, temperature is also lesser compared to the low-lying glaciers.

Herreros et al. (2009) mentioned that there was no change in the quantity of precipitation in the study area during the strong El Niño during 1997–1998, but glacier mass loss was high from our results. However, by using NCEP-NCAR reanalysis data, Pouyaud (2005) could find a correlation between the runoff from the glaciated drainage basins in the Cordillera Blanca and the air temperature. During the strong El Niño seasons, the temperature values were higher in the Coropuna region as well (Figure 5(B)). There exists a high correlation between the glacier mass loss and air temperature in the mid latitude and high-latitude glaciers (Braithwaite 1981) and this explains how the El Niño events were followed by an elevation in annual snowline. The westerly (dry conditions) and the easterly (wet conditions) are the wind anomalies that are not centred over the central part of Altiplano and the location of these wind anomalies are important in determining the spatial pattern of the precipitation anomalies in the central Andes. A study on three drainage basins (Pacific, Titicaca and Amazonas) in Peru shows that the coastal region is having higher rainfall variability and hence higher runoff variability on seasonal and inter-annual timescales (Lavado Casimiro et al. 2012). It is clear that the increase in the runoff, if exists during low precipitation season, is solely due to accelerated glacier ablation.

6. Conclusions

The Nevado Coropuna in the Cordillera Ampato has lost its 26.92% of total glaciated area during the last three decades (from 57.57 to 42.19 km^2) and this decrease in the area followed the pattern of ENSO, particularly when in phase with PDO. This loss of glaciated area is important because the Peruvian Andes contains about 70% all tropical glaciers in the world. Exceptional cases were found during the prolonged El Niño period during 1991–1995, probably due to the influence of the Pinatubo eruption followed by the cooling effect of aerosols in the stratosphere. The fluctuations in area and annual snowline of this icecap in Peru during the phase changes of ENSO is immediate and higher compared to mountain glaciers in Ecuador or Bolivia and this indicates that glaciers in different climate zone (inner and outer tropics, for example) can show different magnitudes of response to identical climatic perturbation (El Niño, for example). Even though a correlation does not exist between the precipitation anomaly and ENSO in this region, the snowline fluctuations followed the ENSO patterns. It is highly recommended to include the influence of spatial barriers (mountain range) in the study of climate variability in the Andean countries.

Acknowledgements

First author acknowledges Fundação de Amparo à Pesquisa do Estado do Rio Grande do Sul (FAPERGS), Brazil, for his PhD research scholarship and Dr. Walter Silverio, University of Geneva, for the nice MMP graph based on meteorological stations near the study site.

Disclosure statement

No potential conflict of interest was reported by the authors.

Funding

This work was supported by the Fundação de Amparo à Pesquisa do Estado do Rio Grande do Sul.

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CAPITULO VIII

Un análisis comparativo del retroceso glaciar en los Andes tropicales usando teledetección

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Investigaciones Geográficas, 51, 3-36, 2016

Un análisis comparativo del retroceso glaciar en los AndesTropicales usando teledetección

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RESUMEN

En este trabajo de investigaciónse analizó el comportamiento de los glaciares de los Andes Tropicales en un clima cambiante; para esto se emplearon imágenes satelitales multiespectrales de diversas fuentes en conjunto con datos meteorológicos. El estudio se enfocó en glaciares representativos de cuatro zonas climáticas diferentes, a saber: trópicos interiores, trópicos exteriores húmedos del norte, trópicos exteriores húmedos del sur y trópicos exteriores secos. Se calcularon los cambios en la línea de nieve máxima anual para el período de 1985 - 2015, y también los cambios decenales en el área entre 1975 y 2015. Adicionalmente, se analizó la tasa de retroceso de los glaciares durante la ocurrencia de El Niño -Oscilación del Sur y la Oscilación Decenal del Pacífico. Se observó que los glaciares tanto de los trópicos interiores como de los trópicos exteriores se sometieron a retroceso durante todo el período de estudio, con énfasis entre 1975 y 1997, lapso coincidente con el período de calentamiento del Pacífico. Se observaron variaciones excepcionales en la altitud de la línea de nieve cuando se produce un evento de El Niño durante la fase cálida de la Oscilación Decadal del Pacífico. Se observó que no hay señales significativas del hiato reciente en el calentamiento global, excepto en los trópicos exteriores secos, localizados cerca de la región subtropical.

Palabras Clave: Andes Tropicales, trópicos interiores, trópicos exteriores, línea de nieve, ENOS, retroceso glaciar.

A comparative analysis of glacier retreat in the Tropical Andes using remote sensing

ABSTRACT

In this research paper, we analysed the behaviour of Tropical Andean glaciers in a changing climate. We used multi-source satellite images as well as meteorological datasets to achieve this objective. Representative glaciers in four different climatic zones, namely the inner tropics, northern wet outer tropics, southern wet outer tropics and dry outer tropics, were considered in this study. Changes in annual maximum snowline during 1985 - 2015 and also the decadal changes in the area between 1975 and 2015 of these glaciers were calculated. Furthermore, we analysed the rate of glacier retreat during the occurrence of El Niño–Southern Oscillation and Pacific Decadal Oscillation. It is observed that the glaciers in both the inner and outer tropics underwent retreat during the study period and most of this retreat occurred during 1975 - 1997 which is parallel with the so-called Pacific shift. Exceptional variations in snowline altitude were observed when an El Niño event occurs during the warm phase of the Pacific Decadal Oscillation. No significant signals of the recent hiatus in global warming were observed, except in the dry outer tropics which are situated near the subtropical region.

Keywords: Tropical Andes, inner tropics, outer tropics, snowline, ENSO, glacier retreat.

Recibido el 25 de mayo de 2016; aceptado el 21 de junio de 2016.

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INTRODUCCIÓN

Los glaciares de montaña de la región tropical son extremadamente sensibles a las fluctuaciones ambientales (HASTENRATH 1994). Las condiciones climáticas en estas regiones favorecen la ablación a lo largo de todo el año en el frente glaciar (FRANCOU *et al*. 2004), generando así una respuesta compleja de estos a las variaciones climáticas (OERLEMANS 1989). El Panel Intergubernamental sobre el Cambio Climático (IPCC) ha reconocido el papel de los glaciares de montaña como indicadores principales del cambio climático (LEMKE *et al*. 2007), debido al rápido tiempo de respuesta, a la sensibilidad a las variaciones climáticas y a los cambios fácilmente observables en su balance de masa. BRADLEY *et al*. (2006) proyectan un aumento de la temperatura en torno a 4°C en altitudes superiores a 4000 m.s.n.m. para el siglo XXI. Estudios recientes han demostrado que los glaciares tropicales de los Andes están retrocediendo a un ritmo alarmante. El estudio del cambio climático en los Andes Tropicales utilizando glaciares como indicadores indirectos del tales variaciones ha sido discutido por muchos investigadores recientemente (VUILLE *et al*. 2008a; JOMELLI *et al*. 2009; VEETTIL *et al*. 2014, 2015, 2016a, 2016b). El retroceso discontinuo de los glaciares andinos tropicales desde la Pequeña Edad de Hielo (PEH) en Ecuador, Perú y Bolivia provee una estimación de las fluctuaciones climáticas desde principios del siglo XVII (VUILLE *et al*. 2008). En un estudio reciente, SCHAUWECKER *et al.* (2014) observaron un aumento en la temperatura del aire de aproximadamente 0,31ºC/década entre 1969 y 1998 y 0,13ºC/década de 1983 y 2012 en la Cordillera Blanca de Perú. Este pequeño descenso en el ritmo de calentamiento en las últimas décadas es más visible cerca de la costa del Pacífico mientras que regiones de gran elevación en los Andes tropicales siguen en aumento (VUILLE *et al*. 2015). Esta observación realizada por VUILLE *et al*. (2015) está asociada al fenómeno conocido como el calentamiento dependiente de la elevación (*Elevation Dependent Warming - EDW*) durante las últimas décadas (PEPIN *et al*. 2015). Un aumento excepcional de la humedad, que es una de las causas de EDW, ha sido observado recientemente en el sur de la Cordillera Blanca por VEETTIL *et al*. (2016b). Se puede argumentar que si el aumento de la humedad en las montañas occidentales relativamente secas es alto, entonces la región húmeda de la cordillera oriental cerca de la cuenca del Amazonas puede ser más susceptible a EDW.

Más del 80% del suministro de agua potable de las poblaciones en las zonas tropicales y subtropicales áridas/semiáridas proviene de los glaciares de montaña (MESSERLI 2001; VUILLE *et al*. 2008a) y más del 95% de todos los glaciares tropicales están situados en América del Sur, cubriendo estos una superficie aproximada a los 2758 km2 (KASER 1999). Durante la estación seca, los países andinos, en especial las zonas más pobladas de Bolivia y Perú, dependen del agua procedente de las cuencas glaciares. Por lo tanto, los glaciares tropicales sirven como amortiguadores críticos de precipitación muy estacional y proporcionan agua para uso doméstico, agrícola o industrial, en ausencia o disminución de las lluvias (VUILLE *et al*. 2008b). Si estos glaciares desapareciesen para siempre, habría consecuencias graves en la disponibilidad de agua para una gran población. Otro problema asociado con el retroceso de los glaciares en los Andes Tropicales, es la formación y la expansión de los lagos glaciares, especialmente en las montañas de las cordilleras Blanca y Oriental localizadas en Perú y Bolivia. La formación y expansión de los lagos glaciares aumenta el riesgo de inundaciones debido a la falta de represas; la posibilidad de tal evento depende de la altura, del área y del volumen de lagos glaciares y de la resistencia natural de las paredes laterales. Deslizamientos, caída de rocas y cascadas de hielo en los lagos glaciares pueden desencadenar vaciamientos abruptos de lagos glaciares (*Glacial Lake Outburst Flood* - GLOF). Por ejemplo, el GLOF de Lago Nº 513 en Chucchún Valle en la Cordillera Blanca en Perú en 2010 se debió a las avalanchas de rocas y hielo cubierto de Nevado Hualcán (VILÍMEK *et al*. 2015). CHEVALLIER *et al*. (2011) han definido tres posibles causas de inundaciones de este tipo en Perú: (1) el aumento del flujo debido a la fusión de los glaciares y el aumento de la presión hidrostática sobre la presa de morrena, (2) hielo y roca que caen en lagos debido a las avalanchas creando así ondas de choque que pueden iniciar desbordamiento y ruptura repentina de las presas/embalses de la morrena, (3) deslizamientos de rocas debido al retroceso de los glaciares. La actividad sísmica que prevalece en las sierras andinas también puede aumentar la probabilidad de inundaciones de este tipo en la región.

En las últimas décadas, el conocimiento de las fluctuaciones glaciales ha mejorado significativamente debido a la comprensión de la relación entre el clima moderno y los glaciares y también debido a la rápida evolución de tecnologías y conceptos en paleoclimatología (SOLOMINA *et al*. 2008). Las imágenes satelitales permiten evidenciar fenómenos asociados al calentamiento global, tales como el aumento del nivel del mar, el retroceso de los glaciares y aceleración de la pérdida de hielo en regiones costeras (BENN & EVANS 1998). El campo de teledetección ha crecido a lo largo de los años en variedad y sofisticación para el monitoreo de la superficie de la Tierra (REES 2006). A veces, la teledetección mediante fotografías aéreas (JORDAN *et al*. 2005) y/o imágenes de satélite es el único método disponible para estudiar los glaciares, sobre todo debido a la falta de acceso a zonas remotas (BOLCH & KAMP 2006). Las imágenes de satélite se pueden utilizar para monitorear la extensión de los glaciares, que es un buen sustituto para el estudio de los cambios de balance de masa para mucho tiempo (HALL *et al*. 1987; PAUL 2000; SILVERIO & JAQUET 2005). La experiencia de incluir/borrar un píxel en clase glaciar decidirá la exactitud de mapeo glaciar (WILLIAMS *et al*. 1997). SURAZAKOV & AIZEN (2006) utilizaron con éxito una combinación de imágenes SRTM (*Shuttle Radar Topographic Mission*) y otros datos topográficos generados a partir de modelos digitales de elevación (MDE) para la estimación de la variación del volumen de los glaciares. La pérdida de masa de los glaciares y la elevación está en relación inversa, cuanto mayor sea la altitud, menor será la pérdida de masa (RABATEL *et al*. 2013).

En escalas interanuales, una fracción significativa de la variabilidad de las lluvias en los Andes Tropicales se relaciona con la aparición de El Niño - Oscilación del Sur (ENOS) (VUILLE & BRADLEY 2000; VEETTIL *et al*. 2014, 2015, 2016a, 2016b), que es una de las fuentes importantes de la variabilidad interanual del clima en la Tierra. Los años de El Niño se asocian generalmente con condiciones calientes y secas, mientras que años de La Niña se asocian con condiciones de frío y humedad en el altiplano andino, sin embargo, las características climáticas de ambos fenómenos no son uniformes en los Andes tropicales (RABATEL *et al*. 2013). Los Andes Tropicales (entre 10ºN y 16ºS) son el conjunto de sitios adecuados para el estudio de la influencia del cambio climático en los glaciares tropicales. Estos glaciares están sujetos a variaciones de temperatura diarias mayores que las variaciones de temperaturas anuales. La precipitación, es una de las variables que determinan el crecimiento y retroceso de los glaciares, especialmente en las regiones tropicales y por lo tanto es importante entender cómo varía la precipitación en la región andina. Es necesario también tener en cuenta que la precipitación cerca de la cuenca del Amazonas es mayor que la precipitación cerca del Pacífico. Se ha encontrado que anomalías de temperatura y precipitación asociadas con ENOS se debilitan hacia el altiplano (GARREAUD 2009). Se ha estudiado igualmente que El Niño - Oscilación del Sur y otros fenómenos globales, como la Oscilación Decadal del Pacífico (ODP) influyen latitudinalmente en el clima andino (GARREAUD *et al.* 2009). La ODP es un índice del clima sobre la base de las variaciones en la temperatura superficial de océano (*Sea Surface Temperature* - SST) en el Pacífico Norte de agua cálida (índice positivo) y regímenes fríos (índice negativo) (MANTUA *et al.* 1997). Estos regímenes pueden persistir por décadas, habiéndose interrumpido tales ciclos desde 1998.

El objetivo de este estudio es comprender la relación entre las interacciones océano-atmosféricas y las fluctuaciones de los glaciares en los Andes Tropicales. Para lo cual fue necesario: (1)calcular las variaciones en la altitud de línea de nieve (ALN) y los cambios decenales en el área de los glaciares andinos en los trópicos interiores y trópicos exteriores mediante teledetección; (2) estimar la teleconexión entre la ocurrencia de El Niño - Oscilación del Sur (ENOS) y las variaciones en las altitudes de la línea de nieve; (3) identificar las anomalías en la temperatura, la precipitación y la humedad a lo largo de los Andes Tropicales; (4) relacionar el comportamiento diferencial, si lo hay, entre los glaciares en los trópicos interiores y trópicos exteriores con la variabilidad del ENOS y las anomalías de temperatura y precipitación; y (5) estimar la influencia de la ODP sobre los efectos de ENOS en las variaciones de la ALN en los Andes Tropicales

ÁREA DE ESTUDIO Y CONDICIONES CLIMÁTICAS

La zona tropical de Sudamérica se puede dividir en dos sub-áreas basándose en las características de precipitación y la ubicación geográfica: los trópicos interiores (Venezuela, Colombia y Ecuador), donde la ablación y acumulación de los glaciares ocurren simultáneamente en todo el año, sin estacionalidad de las precipitaciones (VEETTIL *et al.* 2014); y los trópicos exteriores (Perú, Bolivia y el norte de Chile), donde la estación seca se produce de mayo a septiembre, con condiciones subtropicales y la estación lluviosa de octubre a marzo, con prevalecía de condiciones tropicales (VEETTIL *et al.* 2015, 2016b); en esta última ocurre una acumulación notable sólo durante la temporada de lluvias (JOMELLI *et al.* 2009). SAGREDO & LOWELL (2012) clasificaron los Andes en siete grupos climáticos, con base en el análisis estadístico de tres variables climáticas (temperatura, lluvia y humedad) y en un número seleccionado de glaciares con geometría simple. Tres de los siete grupos definidos en esta clasificación pertenecen a los Andes Tropicales: los trópicos interiores (Venezuela, Colombia y Ecuador), los trópicos exteriores húmedos (cordilleras Occidental Norte y Central de Chile y cordilleras orientales de Perú y Bolivia), y los trópicos exteriores secos (cordilleras occidentales de Bolivia, sur de Perú y norte de Chile) (Fig. 1). En los trópicos interiores, la temperatura y la elevación de la isoterma 0°C varían poco a lo largo del año (KLEIN *et al.* 1999), la acumulación y ablación se producen durante todo el año. Aunque la lluvia se produce durante todo el año, existen dos máximos de precipitación durante marzo-abril y octubre, y un mínimo entre junio y agosto en esa región. La línea de nieve en esta región oscila entre 4500 y 5000 m.s.n.m. Los trópicos exteriores húmedos, donde la cota de nieve se sitúa ~5000 m.s.n.m., se pueden dividir en dos grupos, como trópicos exteriores húmedos del norte (Cordillera Occidental del norte de Perú) y como trópicos exteriores húmedos del sur (Cordillera Oriental de Perú y Bolivia), con base principalmente en características de la precipitación (SAGREDO & LOWELL 2012). La precipitación en los trópicos exteriores húmedos del norte se relaciona principalmente con el cambio de la zona de convección hacia el sur (GARREAUD 2009), mientras que la precipitación en los trópicos exteriores húmedos del sur es más compleja y depende del desplazamiento de la zona de convergencia intertropical (ZCIT) hacia el sur (SAGREDO & LOWELL 2012). Los trópicos exteriores secos se caracterizan por presentar estaciones secas y lluviosas específicas; se produce una acumulación durante el verano austral y los glaciares de la región están sujetos a condiciones extremadamente frías y secas. La línea de nieve en esta región está situada a unos ~5600 m.s.n.m. (NOGAMI 1972).

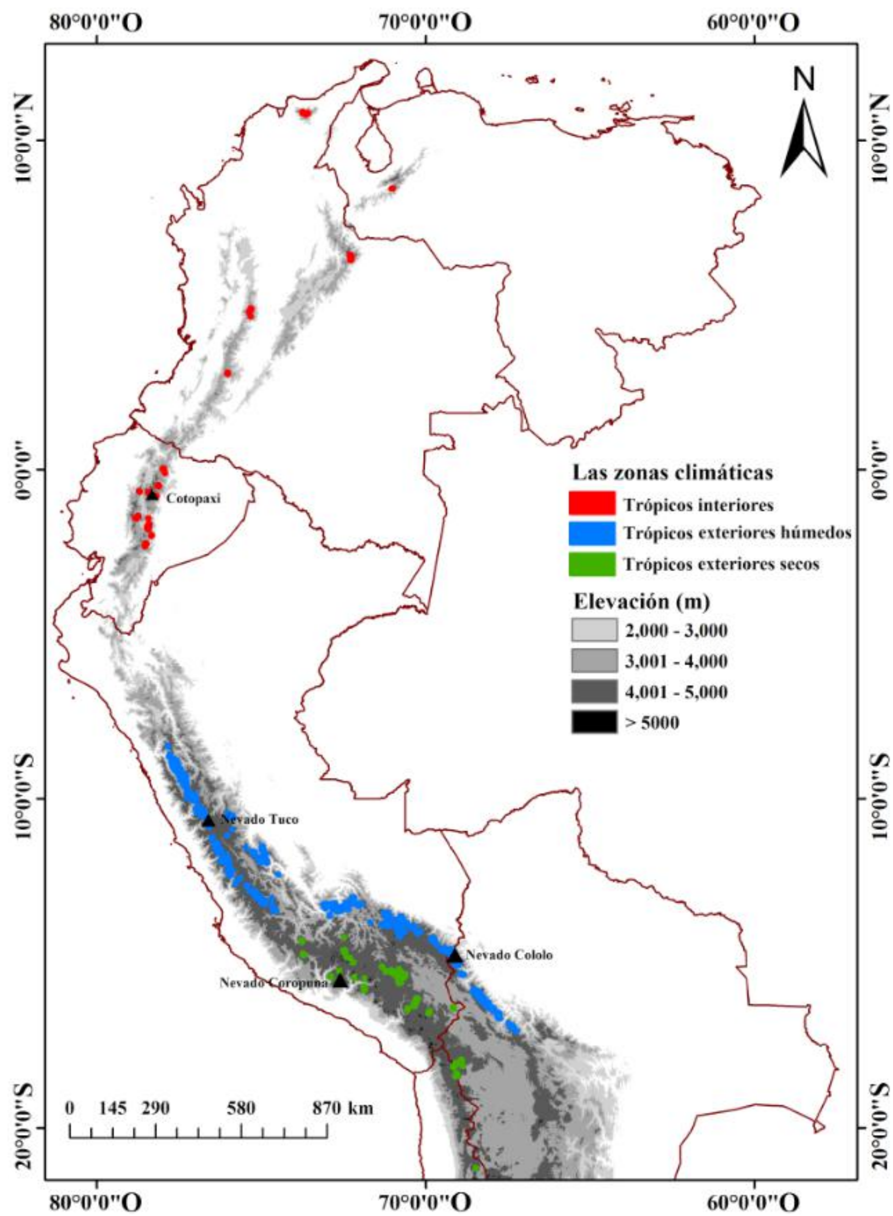

Fig. 1. Distribución de los glaciares en tres grupos de clima en los andes tropicales.

Fig. 1. Glacier distribution in three climate groups in the tropical andes.

Los Andes son la línea divisoria de aguas entre el Océano Pacífico y la cuenca del Amazonas, con una orientación norte-sur (LARAQUE *et al*. 2007). Debido a la ubicación geográfica, teóricamente, los glaciares en la Cordillera Occidental están más influenciados por la circulación atmosférica del Pacífico y en la Cordillera Oriental que por la circulación del Amazonas y el Atlántico. Además, también existe la posibilidad de que los glaciares de los trópicos interiores tengan características de ablación diferentes en comparación con los glaciares de los trópicos exteriores. Para este estudio se consideraron cuatro glaciares de los Andes Tropicales, uno por cada subregión: volcán Cotopaxi (trópicos interiores), Nevado Tuco (trópicos exteriores húmedos del norte), Nevado Coropuna (trópicos exteriores húmedos del sur) y Nevado Cololo (trópicos exteriores secos). Las siguientes sub-secciones describen la ubicación geográfica y las condiciones climáticas de los cuatro sitios de estudio.

Trópicos interiores

Se extienden desde Venezuela hasta Ecuador. En este último país, los glaciares de los trópicos interiores se encuentran en la Cordillera Occidental (0º22’ N - 1º29’S; 78º20’O - 78º48’ O) como en la Cordillera Oriental (0º1’ N - 2º20’S; 77º54’O - 78º33’O). Sobre la Cordillera Oriental, y a unos de 60 km al sureste de la ciudad de Quito, se ubica el volcán Cotopaxi, uno de los volcanes activos más grandes del mundo y es el segundo pico más alto de Ecuador (5897 m.s.n.m.; 0º40 ’S - 78º25’ O; Fig. 2a; 2b).

La Cordillera Oriental está directamente expuesta a los vientos húmedos del este de la cuenca del Amazonas. Alrededor de 20 glaciares irradian hacia fuera del Cotopaxi en todas las direcciones, entre 1976 y 1997, el Cotopaxi había perdido alrededor del 30% de su superficie glaciar (JORDAN *et al*. 2005). Las erupciones más recientes se registraron en 1742-1744, 1768, 1877, 1903-1904 y últimamente en 2015.

La región presenta dos picos de precipitación máxima por año: el primerio entre Marzo y Mayo, y el segundo entre Septiembre y Noviembre. Las variaciones estacionales de las temperaturas no son significativas, pero la variabilidad interanual es considerablemente grande. El viento es el factor principal de estacionalidad en la Cordillera Oriental de Ecuador (FRANCOU *et al*. 2004). En Ecuador, la precipitación es modificada por el sistema montañoso de los Andes, por los fenómenos de mesoescala de los vientos de valle y las corrientes oceánicas (ENOS y Corriente de Humboldt) (BENDIX & LAUER 1992), que afectan en consecuencia, los cambios en el equilibrio de la masa. La temperatura del aire es otro factor que controla el balance de masa y el balance de energía en la región y por lo tanto es más sensible a las variaciones de la temperatura que los trópicos exteriores (FAVIER *et al*. 2004; JOMELLI *et al*. 2009). La incidencia de la radiación de onda corta es máxima cerca del equinoccio (Marzo-Abril a Septiembre) y si no hay precipitación sólida durante este periodo habrá una importante tasa de fusión (RABATEL *et al*. 2013).

Trópicos exteriores húmedos del norte

En esta región de los Andes se ubica la Cordillera Blanca, con una extensión de 180 km de largo y 30 km de ancho (8º30’S - 10º10’S; 77º00’O - 78º00’O). Ubicada en el estado peruano de Ancash, a 400 kilómetros al norte de la capital Lima. Cuenta con 27 picos que alcanzan alturas superiores a los 6000 m.s.n.m. y más de 200 picos que superan los 5000 m.s.n.m. Avalanchas catastróficas recientes ocurrieron en 1962 y 1970 en la Cordillera Blanca. El glaciar del Nevado Tuco (9º55’45’’S; 77º12’18’’O; 5300 m.s.n.m.) se ubica al sur de esta cordillera (Figs. 3a y 3b). Entre otros picos nevados al Nevado del Tuco destacan los nevados Pastoruri, Queullaraju, Jenhuararca, Challwa, Condorjitanca y Santun.

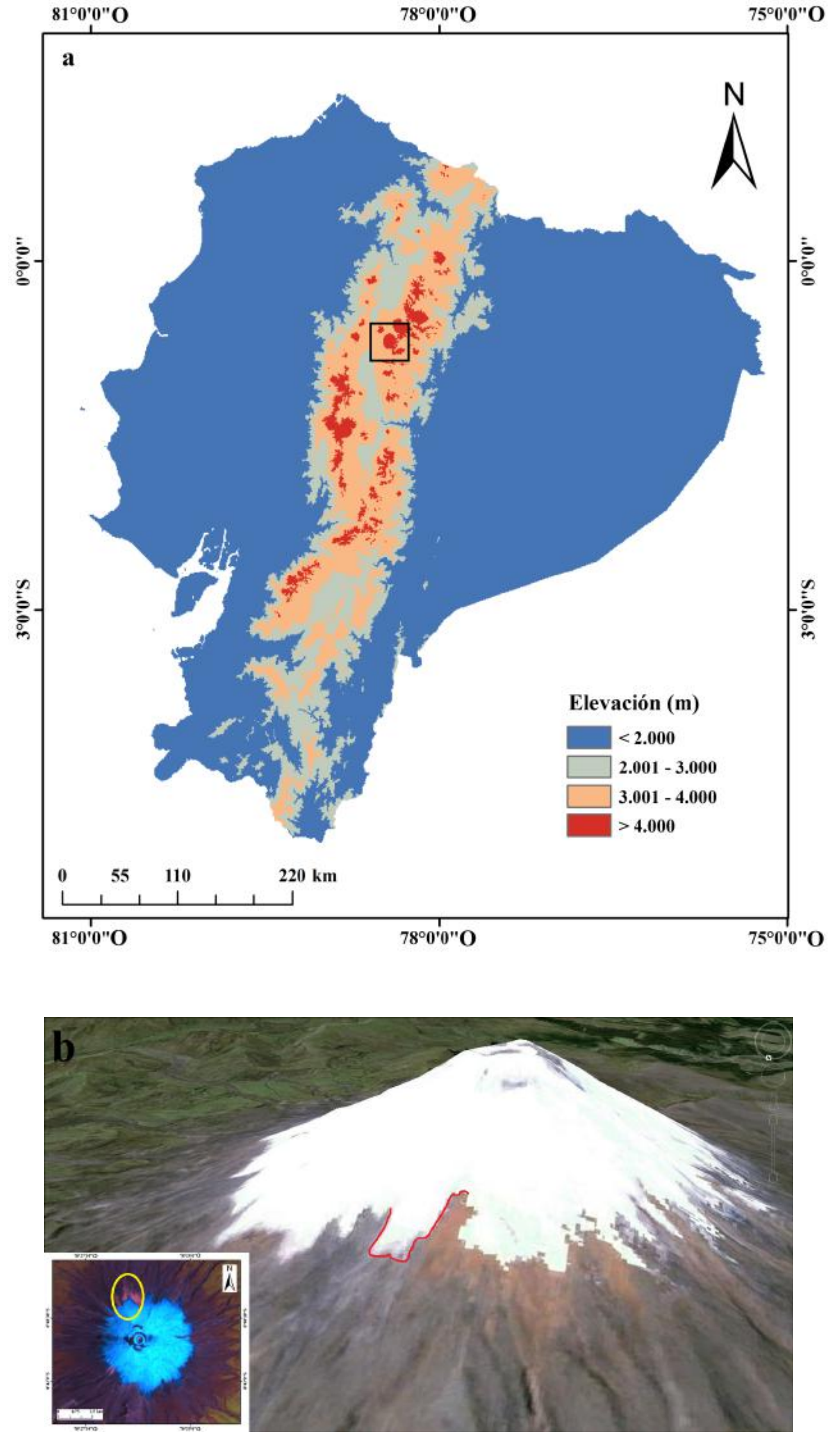

Fig. 2. (a) La ubicación de Cotopaxi y la distribución de la altitud en Ecuador y (b) área de estudio.

Fig. 2. (a) Location of Cotopaxi and altitude distribution in ecuador and (b) Cotopaxi and study area.

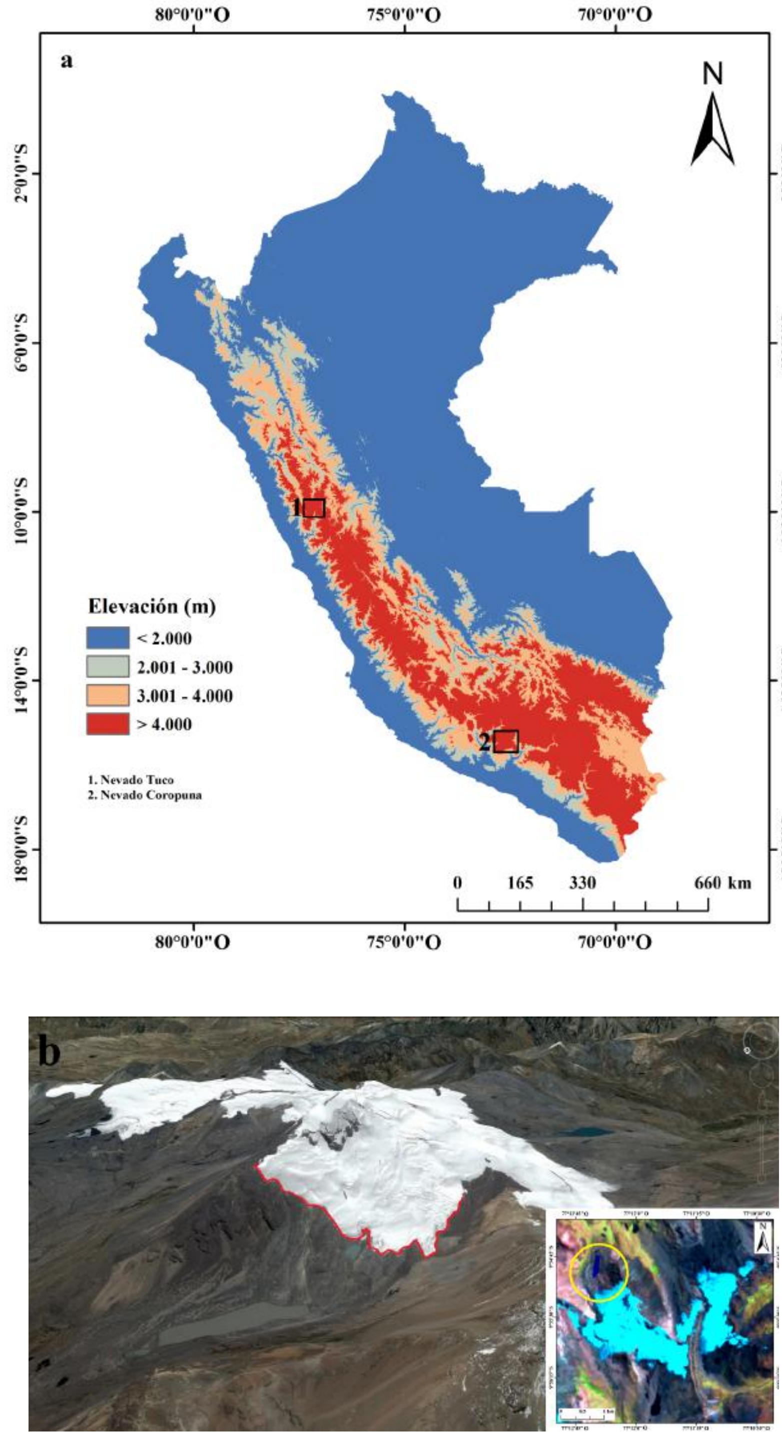

Fig. 3. (a) La localización del Nevado Tuco y Coropuna y distribución de la altitud en Perú (b) área de estudio Nevado Tuco.

Fig. 3. (a) Location of Nevado Tuco and Nevado Coropuna and altitude distribution in Peru (b) Nevado Tuco and study area.

La Cordillera Blanca se caracteriza por una estacionalidad poco acentuada de la temperatura anual (pero con grandes variaciones diarias de temperatura) y el clima alterna entre una estación seca (Mayo a Septiembre) y una estación lluviosa (Octubre a Abril). La mayor (70-80%) parte de la precipitación anual ocurre durante la temporada de lluvias (KASER & GEORGES 1997). Por otro lado, la estación seca en las zonas tropicales de los Andes de Perú se produce durante el invierno austral. A pesar de la proximidad de los glaciares con el Océano Pacífico, los vientos persistentes desde el este determinan la fuente de humedad para la precipitación andina y se derivan principalmente del Atlántico (JOHNSON 1976) y el clima de Perú está fuertemente influenciado por las montañas de los Andes. Aun cuando el ENOS es principalmente un fenómeno de la temperatura superficial del mar en la cuenca del Pacífico ecuatorial, a menudo se reflejen los eventos de calentamiento/enfriamiento en el Atlántico tropical con seis a ocho meses de retraso. Las anomalías de la SST en el Atlántico tropical son sustancialmente más débiles que las observadas en el Pacífico ecuatorial en asociación con ENOS. La acumulación de masa se produce sólo durante la temporada de lluvias, principalmente en las partes superiores de los glaciares, mientras que la ablación se produce durante todo el año. Los glaciares ubicados en las zonas tropicales exteriores y subtropicales se consideran menos sensibles a la temperatura. El Este del Océano Atlántico y la cuenca del Amazonas son las dos fuentes principales de precipitaciones en los Andes Tropicales, principalmente debido a los vientos estacionales del Este (VUILLE & KEIMIG 2004). Las tasas de mayor precipitación se presentan con mayor frecuencia en las laderas Este, probablemente debido al aumento del transporte de la humedad de la cuenca del Amazonas. Durante los eventos del ENSO 1982 - 1983 y 1992 se observaron disminuciones en las tasas de precipitación; no obstante, el ENOS de 1997 no afectó las tasas de precipitación (HERREROS *et al*. 2009). Hay 15 estaciones meteorológicas operadas por el SENAMHI (Servicio Nacional de Meteorología e Hidrología del Perú) distantes a aproximadamente 60 km del lugar de estudio. Desafortunadamente, la mayoría de las estaciones meteorológicas e hidrológicas han dejado de funcionar o tienen datos incompletos.

Trópicos exteriores húmedos del sur

Los glaciares bolivianos se encuentran en la región definida como trópicos exteriores húmedos del sur, y se caracterizan por la baja variabilidad de la temperatura, alta afluencia de radiación solar durante todo el año, alta variabilidad de la humedad y la precipitación estacional (RABATEL *et al*. 2012). El Nevado Cololo (14°50'S; 69°06'O; 5859 m.s.n.m.) de la Cordillera de Apolobamba se encuentra en esta región (Figs. 4a y 4b). Desde una perspectiva glaciológica, existen tres tipos de cambios de ablación (RABATEL *et al*. 2012): (1) los de tasa de fusión más alta debido a la radiación solar (Octubre - Diciembre), (2) los de tasa de ablación mayor debido a la fusión (Enero - Abril) y (3) los de tasa de ablación limitada debido a la pérdida de energía por radiación de onda larga (Mayo - Agosto). Teniendo en cuenta que si la nieve se produce entre Mayo y Agosto pueden permanecer durante la estación seca (RABATEL *et al*. 2012), esto se suma las dificultades para asignar el frente glaciar usando sensores remotos.

En Bolivia el clima varía desde el tropical al clima desértico frío, dependiendo de la altitud (SEILER *et al*. 2013a, 2013b). La temperatura media anual oscila entre 0° y 30°C y rangos de precipitación de 300 mm a 3000 mm por año. La precipitación y su variabilidad interanual están vinculadas a las anomalías de las SST tropicales y las circulaciones atmosféricas (ARNAUD *et al*. 2001; VUILLE 1999). Las lluvias se presentan en los meses de Diciembre a Marzo. El verano austral (Diciembre -

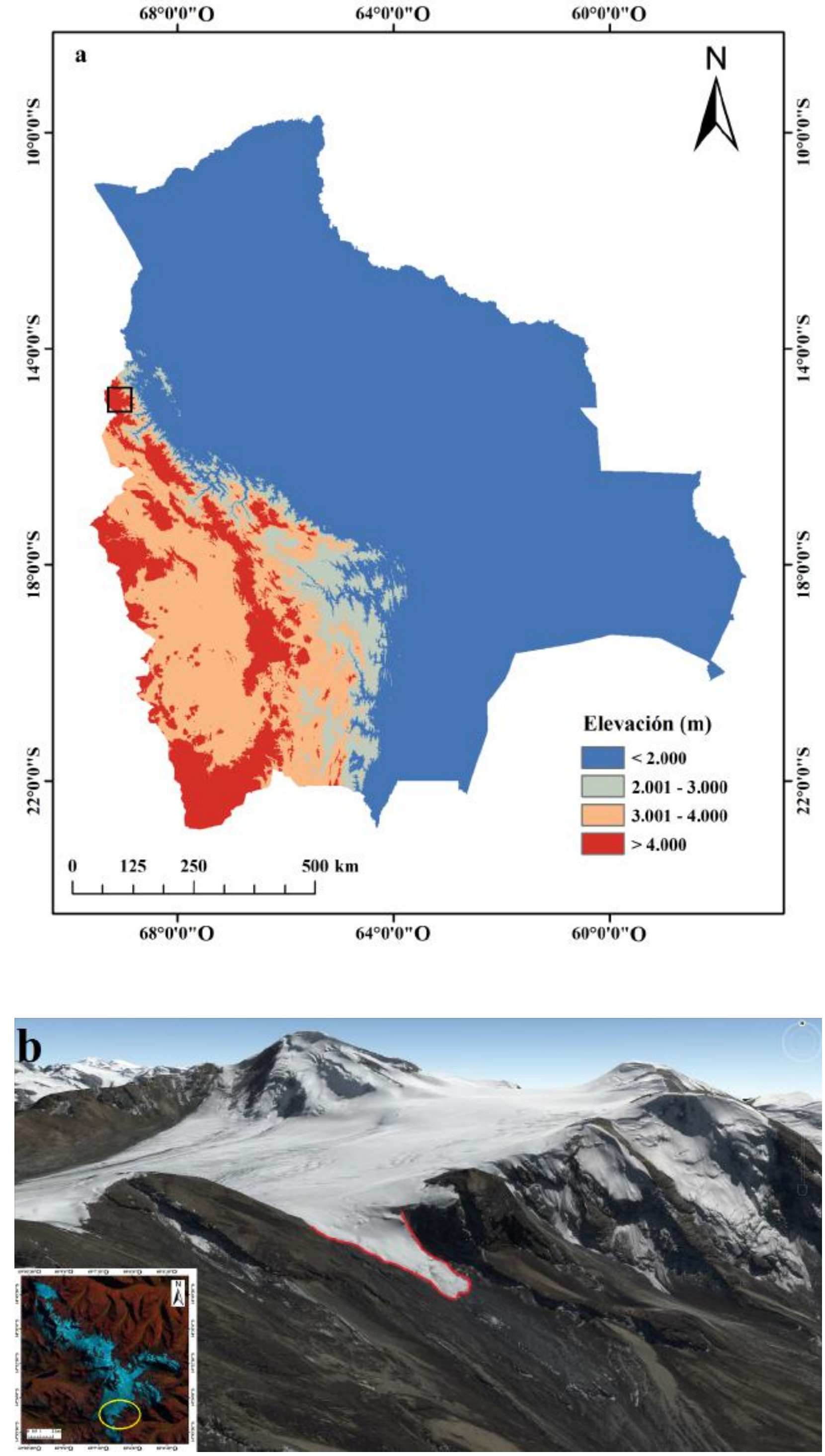

Fig. 4. (a) La localización del Nevado Cololo y distribución de la altitud en Bolivia y (b) área de estudio.

Fig. 4. (a) Location of Nevado Cololo and altitude distribution in Bolivia and (b) Nevado Cololo and study area.

Enero - Febrero - DEF) se caracteriza por un sistema de baja presión que aumenta los vientos del este (vientos comerciales) que transportan la humedad desde el Atlántico tropical (norte) hacia el continente. Este contenido de humedad es desviado por los Andes y se transporta hacia el sur causando un aumento de las lluvias en el océano Atlántico (MARENGO *et al*. 2004). Con la liberación de calor de condensación en el Amazonas se forma un sistema de alta presión en Bolivia, causado por las laderas de los Andes. Este sistema hace que se transporte humedad desde el Amazonas a las tierras altas de Bolivia (SEILER *et al*. 2013a; VUILLE 1999). En el invierno austral (Junio – Julio - Agosto - JJA), hay menos transporte de humedad; este se produce desde el Atlántico norte tropical hacia el continente y los frentes fríos provenientes del Polo Sur penetran en las tierras bajas de Bolivia, disminuyendo así la temperatura y precipitación (GARREAUD 2009). Los vientos del oeste que prevalecen en Bolivia impiden el transporte de humedad de los Andes durante el invierno austral (VUILLE 1999). Las tres principales fuentes de variabilidad climática en Bolivia son: (1) ODP, (2) ENOS y (3) oscilación antártica (OA) (SEILER *et al*. 2013b). Sobre la base de las observaciones meteorológicas, SEILER *et al*. (2013b) se afirmaron que el clima de Bolivia se está calentando a un ritmo 0,1^{o}C cada década y sigue las normas ODP.

Trópicos exteriores secos

La Zona Volcánica Central (ZVC) de la Cordillera Ampato, al sur de Perú, forma parte de esta región de los Andes Tropicales. La Cordillera Ampato se compone de 93 glaciares, con un espesor medio de 40 m y una superficie total de 146,73 km2 (basado en fotografías aéreas de 1962). El Coropuna (15^{o}24'-15^{o}51'S; 71º51'-73º00'O; Figs. 3a y 5), con 6426 m.s.n.m., es el pico más alto de esta cordillera y el volcán más alto de Perú (RACOVITEANU *et al*. 2007). Muchas personas en la parte norte de la ciudad de Arequipa dependen del deshielo de sus glaciares para el suministro de agua. El retroceso de los glaciares de la Cordillera de los Andes de Perú comenzó a mediados de 1980 (SALZMANN *et al*. 2013). RACOVITEANU *et al*. (2007) encontraron que el tamaño del Coropuna se redujo de 82,6 km2 en 1962, a 60,8 km2 en 2000. El balance de masa glacial en esta región es muy dependiente de las variaciones en la precipitación (WAGNON *et al*. 1999).

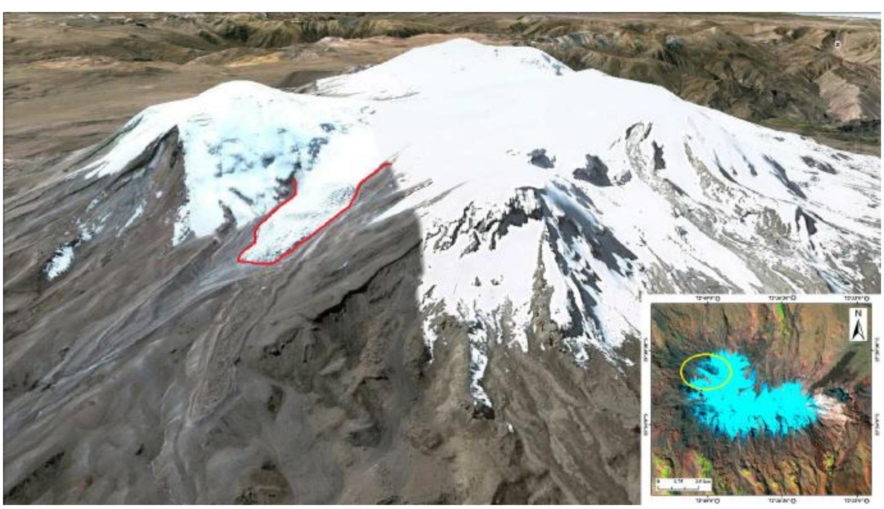

Fig. 5. El Nevado Coropuna y su área de estudio.

Fig. 5. Nevado Coropuna and study area.

Las precipitaciones en la región del Coropuna dependen principalmente de la circulación de las masas de aire desde el Océano Atlántico hacia el oriente tropical (HERREROS *et al*. 2009). Sin embargo, en el Pacífico, los patrones de circulación de aire son importantes en la determinación del clima en esta región. Al igual que los glaciares de las regiones de trópicos exteriores y subtrópicos, el Coropuna tiende también tener una línea de equilibrio (LE) por encima de la isoterma de 0°C, mientras que en las zonas tropicales interiores están cerca de la isoterma de 0°C. Por lo tanto, los glaciares ubicados en las zonas tropicales exteriores y subtropicales se consideran insensibles a la temperatura (KASER 1999). El Este del Océano Atlántico y la cuenca del Amazonas controlan la precipitación en los Andes Tropicales, principalmente por los vientos estacionales del este (VUILLE & KEIMIG 2004). La variación estacional de la temperatura es pequeña, mientras que la precipitación es mayor; 70% - 90% de esta precipitación ocurre durante el verano austral (Diciembre - Marzo). La estación seca en las zonas tropicales de los Andes de Perú se produce durante el invierno austral. Las tasas de precipitación más altas tienen lugar en las laderas orientales, probablemente debido al aumento del transporte de la humedad de la cuenca del Amazonas.

MÉTODOS

Se emplearon datos tanto de teledetección como meteorológicos. Los datos de teledetección incluyen las imágenes multiespectrales de varios sensores en el período 1984 - 2014 y modelos digitales de elevación (MDE). Los datos meteorológicos incluyen la precipitación y la temperatura superficial del aire durante el período de estudio. Índices del ENOS y ODP también están incluidos en este último conjunto de datos.

En este estudio fueron utilizadas imágenes de satélite de diversas fuentes. Las imágenes disponibles de Landsat son ortorrectificadas, por lo tanto, no fue necesaria ninguna corrección geométrica durante su pre-procesamiento. Estas fueron descargadas en formato GeoTIFF desde el portal en línea del USGS (https://www.usgs.gov/). Las imágenes EO—1 ALI fueron obtenidas a través del mismo portal del USGS y las imágenes IRS LISS III (en el año 2012, a causa de la interrupción de las imágenes Landsat) fueron obtenidas del INPE (Instituto Nacional de Pesquisas Espaciais; http://www.inpe.br/) de forma gratuita. Finalmente, las imágenes ASTER fueron obtenidas de la Reverb de la NASA (http://reverb.echo.nasa.gov/reverb/) luego de registrar el proyecto. Además de las imágenes multiespectrales, también se utilizaron modelos digitales de elevación (MDE) generados a partir de ASTER GDEMpara calcular la ALN, con una resolución espacial de 30 m. Las morrenas de los glaciares están cubiertas de nieve durante la temporada alta de la pluviosidad y la ablación es mayor durante la temporada de escasez de precipitaciones. Es difícil calcular el área o la ALN del glaciar a partir de imágenes de satélite durante temporadas de mayor precipitación, por lo tanto, todas las imágenes utilizadas en este estudio corresponden al invierno austral. Se hicieron corregistros de imágenes multiespectrales y MDE, luego fueron calibradas radiométricamente antes de la aplicación de los algoritmos de procesamiento de imágenes. Todas las imágenes fueron corregidas con el ángulo cenital solar. El procesamiento de las imágenes se realizó utilizando Erdas Imagine y ESRI ArcGIS 10.1.

El Índice Oceánico de El Niño (ION) es uno de los índices primarios utilizados para monitorear el fenómeno de ENOS. El ION se calcula como las anomalías de temperatura promedio de la superficie del mar en una zona del océano al Este - centro del Pacífico ecuatorial, que se llama región del Niño - 3.4 (5ºN - 5ºS; 170ºO - 120ºO). Además, el tiempo promedio de tres meses se calcula con el fin de aislar mejor la relación con la variación del fenómeno ENOS. Los valores del ION fueron descargados del sitio web del Centro de Predicción del Clima (CPC), de la Administración Nacional Oceánica y Atmosférica (NOAA) (http://www.cpc.ncep.noaa.gov/). Dentro de los valores de la región Niño - 3.4 utilizados en este estudio, los episodios fríos y calientes se definen cuando se mantiene un umbral del +/- 0,5°C durante un mínimo de cinco meses consecutivos. Se sabe que la temperatura media de la superficie del mar aumentó entre el período de 1975 a 2008. La ODP es un índice basado en la variación de la SST en el Pacífico norte; sus registros fueron descargados del sitio web la Agencia Meteorológica de Japón (http://www.jma.go.jp/jma/indexe.html). Los valores predeterminados por el índice del ODP se derivan de las anomalías de SST mensuales en el norte del océano Pacífico desde 20ºN al Polo Norte. A diferencia de ENOS, los eventos del ODP pueden persistir durante varias décadas.

Los procedimientos se dividieron en dos etapas; en la primera etapa, las imágenes de satélite se utilizaron para calcular la altitud límite de las nieves (ALN) durante la estación seca. Esta ALN se puede utilizar como una estimación de la altura de la línea de equilibrio del año. En la segunda etapa, los cambios en la ALN fueron contrastados con la ocurrencia de fenómenos ENOS y ODP. Las imágenes libres de nubes adquiridas durante los últimos meses de verano, se utilizaron para evitar la dificultad de definir el margen de hielo debido a la ablación excesiva. Los canales visibles e infrarrojos se utilizan para trazar efectivamente la línea de nieve en condiciones deseadas (ARNAUD *et al*. 2001). La identificación de la nieve y del hielo utilizando la teledetección, en la teoría es fácil, pero en la práctica no es tan sencillo (ALBERT 2002). Esto se debe a la reflectancia espectral de nieve y hielo (Fig. 6) varía en función de la cantidad de impurezas, a la película de agua debido a la fusión del hielo, y a los cambios con el envejecimiento. Dos sistemas de satélites utilizados en las últimas dos - tres décadas, Landsat y SPOT, sufrieron limitaciones de resolución espacial y espectral, respectivamente. Una de las desventajas de uso de imágenes de satélite es que es posible que tengan algoritmos específicos para las imágenes de un lugar

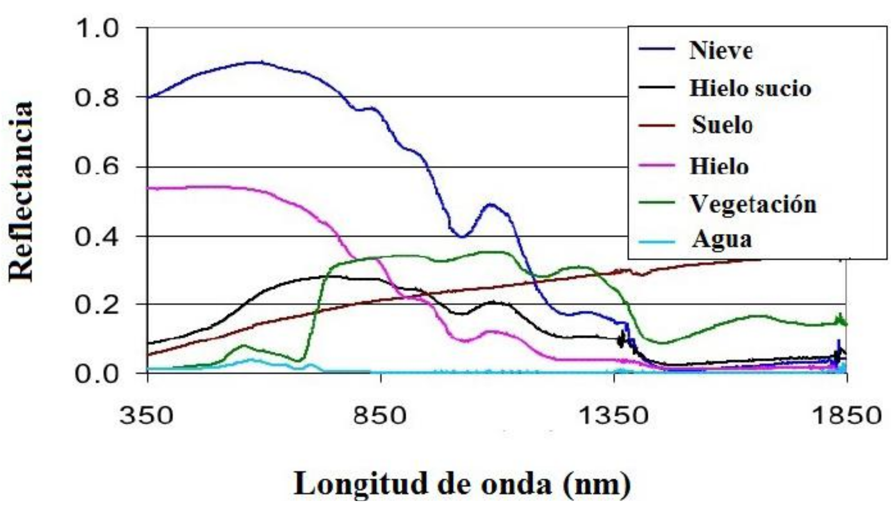

Fig. 6. Curvas de reflectancia espectral de hielo y nieve (fuente: KULKARNI *et al*. 2007).

Fig. 6. Spectral reflectance curves of ice and snow (source: KULKARNI *et al*. 2007).

a otro o de una estación a otra. Con el fin de garantizar la precisión y reducir los errores, las imágenes fueron corregidas atmosféricamente y fueron corregistradas antes del procesamiento adicional.

Uno de los métodos más eficaces para el mapeo de los glaciares es la delimitación manual, pero en este caso no se puede hacer porque un gran número de imágenes de más de 30 años fueron usadas. La superficie de los glaciares se puede calcular mediante la aplicación de un umbral para las imágenes de los índices de nieve por diferencia normalizada (NDSI). Las imágenes NDSI por lo general se calculan a partir de las bandas del verde (TM2, 0,52 - 0,60 μm) y del infrarrojo medio (TM5, 1,55 - 1,75 μm) de Landsat, mediante la siguiente ecuación:

$$NDSI = [(TM2-TM5) / (TM2 + TM5)] \qquad \text{Ec. 1}$$

El valor del umbral que se aplica al delinear el borde del glaciar puede variar de un lugar a otro e incluso de imagen en imagen. Métodos de NDSI han sido ampliamente utilizados por muchos investigadores en la Cordillera Blanca (SILVERIO & JAQUET 2005; RACOVITEANU *et al*. 2008).

Los cambios en el área no siempre son un buen indicador de las variaciones anuales del clima debido a las variaciones interanuales en la cubierta de nieve. Sin embargo, la ALN es un buen indicador de las variaciones climáticas anuales y se puede calcular siguiendo lo propuesto por RABATEL *et al*. (2012). La mayor ALN calculada al final de la estación seca (Mayo - Agosto) puede ser tomada como representativa de la LE, especialmente en los trópicos exteriores y subtrópicos. La comparación y validación de los valores calculados de ALN con los datos de campo del glaciar Zongo en Bolivia y del glaciar Artesonraju en Perú, indican que la ALN representa adecuadamente la LE y por lo tanto se puede utilizar para medir los cambios anuales de balance de masa (RABATEL *et al*. 2012). En contraste con los trópicos interiores, hay una fuerte estacionalidad de la precipitación en los trópicos externos y debido a esta característica, se puede utilizar el valor más alto de ALN durante la estación seca como representante de LE del año. Para el cálculo de la ALN a partir de imágenes de satélite Landsat, fue utilizada la combinación de bandas espectrales de falso colores 5-4-2 con un umbral aplicado a TM2 y TM4. Con el fin de obtener la ALN con la precisión requerida, el umbral aplicado a TM4 varía entre 60 y 135, y el aplicado a TM2, oscila entre 80 y 160, teniendo en cuenta que las imágenes con una resolución radiométrica de 16 bits se convierten en 8 bits para aplicar el mismo algoritmo. Los ejemplos de diversos métodos utilizados para calcular la altitud línea de nieve se muestran en la figura 7.

El análisis de series de tiempo permite entender mejor la teleconexión entre ocurrencias de ENOS (y ODP) y las variaciones de la línea de nieve. Cabe señalar que existe un retraso de 1 a 3 meses entre la ocurrencia de El Niño/La Niña y sus variaciones en las variables meteorológicas del continente, en función de la distancia geográfica de la costa del Pacífico. En cada temporada de lluvias se inicia la línea de nieve, pero durante la estación seca la ALN no es estable en relación con su posición al final de la estación lluviosa, lo cual debe ser tenido en cuenta incluso si no se presenta ablación por fusión debido a las bajas temperaturas de la estación seca; o si la sublimación todavía está presente y es aún mayor que en la época de lluvias debido a las condiciones del cielo con pocas nubes y la presencia de una alta gradiente vertical de humedad. Los cambios en la ALN en cada año deben ser contrastados con la ocurrencia de las fases caliente y frío del ENOS y los regímenes caliente y frío de la ODP.

Errores en el cálculo de ALN utilizando imágenes multiespectrales

Hay algunas posibilidades de errores difíciles de estimar al calcular la ALN.

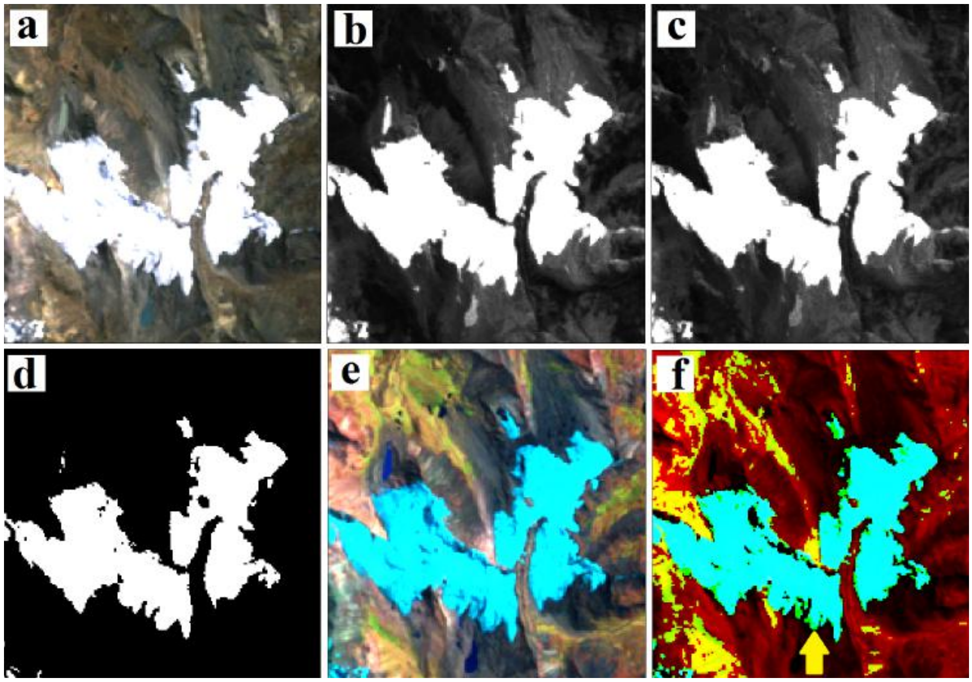

Fig. 7. Diferentes combinaciones de bandas y relaciones de bandas aplicadas en la imagen landsat-5 adquirida el 14 de agosto de 1991 para identificar de la línea de nieve: (a) la combinación de bandas espectrales 3, 2 y 1; (b) relación de tm3/tm5; (c) relación de tm4/tm5; (d) el índice de diferencia normalizada da nieve (ndsi) con el umbral 0,6; (e) la combinación de bandas espectrales 5, 4 y 2; (f) igual a (e) con el umbral 120 y 135 para las bandas de 4 y 2, respectivamente y la flecha amarilla indica la posición de la línea de nieve en el nevado tuco.

Fig. 7. Different band combinations and band ratios applied to a landsat-5 image aquired on 14 august 1999 to identify the snowline: (a) an image with 3-2-1 band combination; (b) tm3/tm5 ratio image; (c) tm4/tm5 ration image; (d) normalized difference snow index (ndsi) image after applying a threshold of 0.6; (e) an image with 5-4-2 band combination; (f) same as (e) with a threshold of 120 and 135 for the bands 4 and 2 respectievly and the yellow arrow shows the position of the snowline of nevado tuco.

Debido a las temperaturas muy bajas en los sitios de estudio, la presencia de hielo limpio es rara, o este a menudo está cubierto de nieve. Esto evita la detección de la zona de ablación en la parte inferior, que es bastante inusual para una lengua glaciar clásica. La ALN calculada también depende en gran medida de la resolución del MDE utilizado. En el corregistro de todas las imágenes y el MDE fue necesario calcular las ALN debido a la topografía extremadamente escarpada y de aspecto variable en algunas partes de los sitios de estudio. Estas fueron calculadas, sólo para áreas específicas de los glaciares, y no a partir de toda su extensión debido a la variabilidad antes mencionada. Las áreas de estudio fueron seleccionadas en pendientes suaves y en el invierno cuando se produce mayor iluminación solar. El error en el cálculo de las ALN dependió del error en el corregistro de las imágenes, de la resolución horizontal y vertical de la MDE y de la pendiente local. En una parcela plana, un error horizontal no tendría ningún impacto en la determinación de la altitud pero con una inclinación de 45°, un error horizontal tendría la misma magnitud del error de altitud. Los errores verticales resultantes, sin embargo, estuvieron dentro de la precisión requerida para este trabajo. Esto es porque todos los glaciares considerados en este estudio tienen poca pendiente (<20°)

y la precisión vertical de ASTER GDEM (V2) está dentro del ámbito de aplicación de este valor. Una de las dificultades en la estimación de la línea de equilibrio sobre la base de ALN en los trópicos interiores (Ecuador), es el hecho de que hay dos estaciones de alta precipitación por año, y por lo tanto no aparece como eficaz. Por lo tanto, se utilizó la línea de nieve máxima calculada durante la temporada de bajas precipitaciones.

RESULTADOS

Los cambios decadales en el área y las variaciones anuales de la línea de nieve son una buena combinación para entender la influencia de las variaciones climáticas en los glaciares andinos en los trópicos. En algunos casos, se excluyeron resultados debido a la falta de imágenes, el exceso de nieve o la presencia de nubosidad.

Para el análisis de las series de tiempo, los índices ENOS y ODP fueron representados gráficamente. Con el fin de entender cómo cambian las tres variables meteorológicas durante el período de estudio, incluimos los resultados de cuatro estudios recientes (VEETTIL *et al*. 2014, 2015, 2016a, 2016b). La discusión sobre las anomalías en las variables meteorológicas (precipitación, temperatura del aire y la humedad) se basaron en los resultados (VEETTIL *et al*. 2014, 2015, 2016a, 2016b). La figura 8 muestra la distribución de tendencia de precipitación (mm/década), temperatura (ºC/década) y humedad relativa (%/década) durante el periodo de 1901-2012.

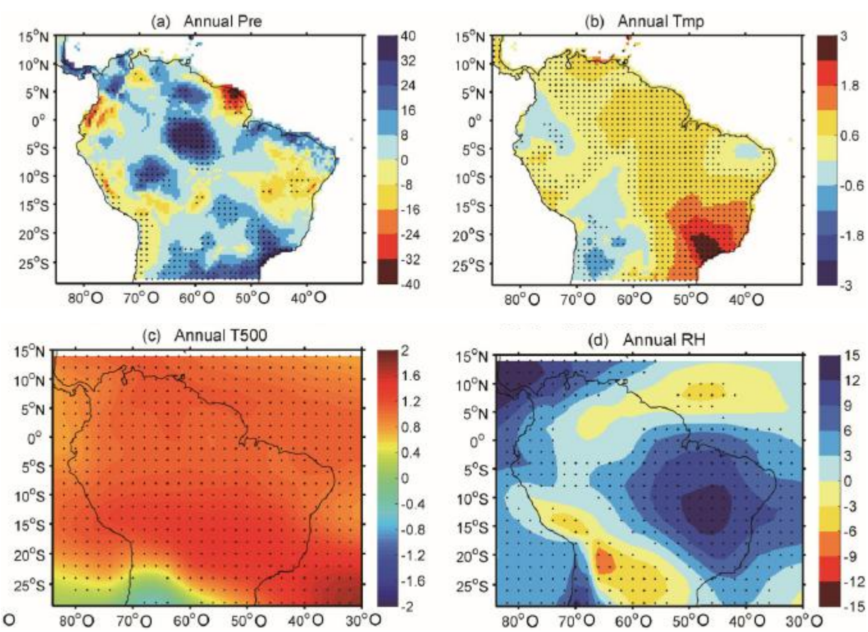

Fig. 8. La distribución de la tendencia de (a) precipitación; (b) temperatura; (c) temperatura a 500 hpa y (d) humedad relativa (a y b se basan en el conjunto de datos cru; c y d se basan de las medias mensuales de noaa-cires reanálisis-v2. Aquí se utilizó la prueba de mann-kendall para detectar tendencias y puntos denotamos que las tendencias son significativas con un nivel de confianza superior al 95%).

Fig. 8. Trend distribution of (a) precipitation; (b) temperature; (c) temperature at 500 hpa and (d) relative humidity (a and b are based on cru datasets; c and d are based on noaa-cires reanalysis-v2 monthly averages. Here we used the mann-kendall test for detecting trends and dots denote the trends that are significant with a confidence level higher than 95%).

Cambios decadales en el área de los glaciares

Una recopilación de los cambios del área del glaciar considerada en este estudio durante las últimas décadas se muestra en la Tabla 1. Esto incluye los resultados de otros estudios. El área se calculó sobre la base de NDSI y de la relación de las bandas.

Glaciar	Año	La imagen usada	Referencia	Área (km2)
Cotopaxi	1976	fotografías aéreas	JORDAN *et al*. 2005	19,24
	1987	Landsat TM	ESTE ESTUDIO	13,65
	1997	fotografías aéreas	JORDAN *et al*. 2005	13,45
	2001	Landsat ETM+	VEETTIL 2012	12,76
	2014	Landsat OLI	ESTE ESTUDIO	11,37
Nevado Tuco	1975	Landsat MSS	ESTE ESTUDIO	4,33
	1987	Landsat TM	ESTE ESTUDIO	3,32
	1998	Landsat TM	ESTE ESTUDIO	2,85
	2005	Landsat TM	ESTE ESTUDIO	2,58
	2015	Landsat OLI	ESTE ESTUDIO	2,21
Nevado Cololo	1975	Landsat MSS	ESTE ESTUDIO	43,7
	1989	Landsat TM	SANCHES 2013	35,44
	1997	Landsat TM	SANCHES 2013	31,4
	2008	Landsat TM	SANCHES 2013	25,51
	2015	Landsat OLI	ESTE ESTUDIO	17,84
Nevado Coropuna	1975	Landsat MSS	ESTE ESTUDIO	66
	1986	Landsat TM	VEETTIL *et al*. 2016a	57,57
	1997	Landsat TM	VEETTIL *et al*. 2016a	52,51
	2006	Landsat TM	VEETTIL *et al*. 2016a	48,63
	2015	Landsat OLI	ESTE ESTUDIO	43,04

Tabla 1: Cambios decenales en el área de los glaciares entre 1975 y 2015.

Table 1: Decadal changes in the area of glaciers between 1975 and 2015.

El porcentaje de pérdida de superficie de los glaciares entre 1975 y 2015 en Cotopaxi, nevado Tuco, Nevado Cololo y Nevado Coropuna fueron 57,65%, 48,96%, 59,17% y 34.78%, respectivamente. Es interesante observar que la mayor tasa de retroceso glaciar ocurrió durante 1975 - 1998 en todas las regiones de estudio (30,1%, 34,18%, 28,14% y 20,43%, respectivamente), especialmente en la Cordillera Blanca.

Variaciones en la ALN durante el periodo 1985 - 2015

Debido a la variabilidad de la extensión de la capa de nieve, se puede utilizar la ALN como un indicador más preciso de tiempo en lugar del área del glaciar. En Ecuador, la variabilidad interanual del balance de masa es controlada por las variaciones anuales de la temperatura del aire y la tasa media de la ablación es sustancialmente constante a lo largo del año en escalas de tiempo estacionales. Esto creó algunas dificultades en la estimación de la ALN del Cotopaxi debido al exceso de la cubierta de nieve, por lo cual se omitieron algunos de los resultados. En un estudio basado en la cubierta de hielo del volcán Sajama en Bolivia (trópicos exteriores secos), VEETTIL *et al*. (2013) correlacionaron los cambios de área de los glaciares con ENOS. Sin embargo, la correlación con la variación de la zona no mostró resultados aceptables, y el índice ODP no se consideró al interpretar la influencia de ENOS en el glaciar. Variaciones en la ALN del Nevado Coropuna, que se encuentra en condiciones climáticas similares, se presentan en este estudio. Las cotas de nieve calculadas en los cuatro sitios de estudio se presentan gráficamente en la figura 9.

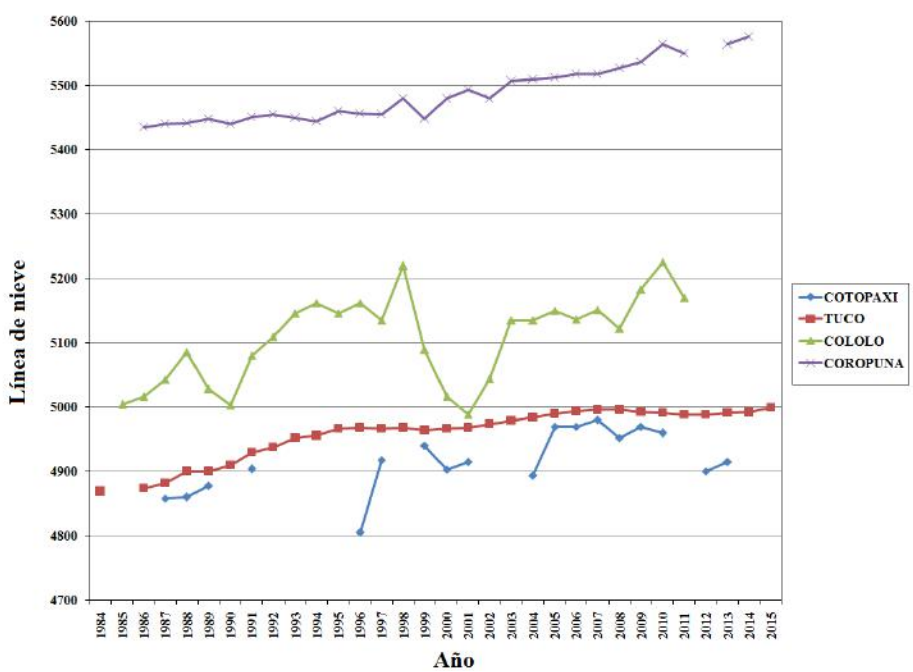

Fig. 9. Variaciones en la línea de nieve durante 1985 - 2015.

Fig. 9. Variations in snowline during 1985 - 2015.

Hubo una tendencia creciente en la línea de nieve durante este período y sus fluctuaciones se han perturbado con los cambios de fase de ENOS y ODP. En todas las áreas de estudio hubo un ligero descenso de la ALN hacia el final del siglo XX. Para el caso particular del Nevado Cololo, a pesar de sus altos índices de precipitación y la posición elevada de su ALN en comparación con el Cotopaxi y Tuco, la velocidad de retroceso de su glaciar fue mayor (160 m entre 1985 y 2011). Esto podría deberse a una de dos causas: 1) los glaciares de los trópicos exteriores húmedos están sujetos a mayores fluctuaciones en la ALN en comparación con los glaciares en Ecuador, o 2) la ausencia de una temporada de precipitaciones definida provocó errores en el cálculo de ALN en el caso de los glaciares de Ecuador. Teniendo en cuenta el mayor valor de ALN y su variación fueron consideradas aquí como representativos de LE, puede que esta no ocurra al final del año hidrológico y que en caso de producirse, dependa de nevadas estacionales. En este estudio, se consideraron dos parámetros geométricos (altitud del glaciar y su exposición), antes de hacer una conclusión sobre la velocidad de retroceso de los glaciares.

Los índices de ENOS y ODP

Los índices de ENOS (se usó el Índice Oceánico de El Niño - ION) y ODP están disponibles hasta la fecha. A partir de estos se creó la serie de tiempo utilizando los datos de la NOAA y del Servicio Meteorológico de Japón, respectivamente (Fig. 10).

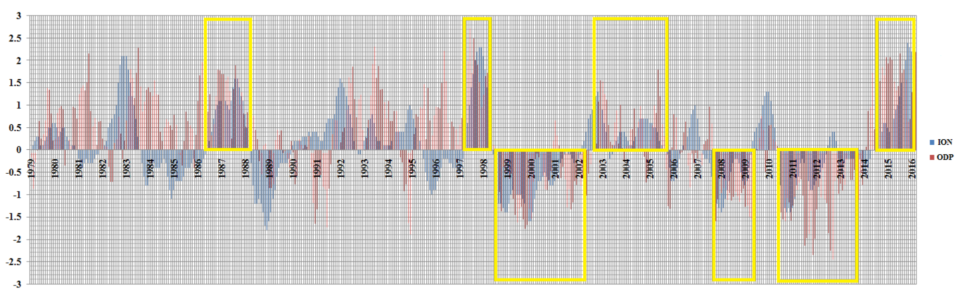

Fig. 10. Los índices de ion e odp entre 1979 y 2016. Cajas amarillas muestran cuando hay una ocurrencia combinado de fases similares de ambos.

Fig. 10. Oni and pdo indices between 1979 and 2016. Yellow boxes show when similar phases of both occur together.

DISCUSIÓN

Los glaciares en los trópicos tienen dos características especiales: estos son sometidos a mayores niveles de energía de forzamiento debido a la ubicación latitudinal y altitudinal específica, y la acumulación y la ablación se produce simultáneamente (este equilibrio puede ser roto por los desequilibrios climáticos) debido a la precipitación durante todo el año (CHEVALLIER *et al*. 2011; KASER & OSMASTON 2002). Las influencias de ENOS y otros fenómenos océano - atmosféricos en el clima andino varían a lo largo de su extensión (GARREAUD 2009). En los trópicos exteriores, la distribución anual de la precipitación, particularmente durante Diciembre - Febrero, influye en la tasa de fusión anual (FAVIER *et al*. 2004). Sin embargo, las variaciones de precipitación toman un tiempo más largo para afectar el frente del glaciar en comparación con las fluctuaciones de temperatura (BONNANO *et al.* 2013).Los glaciares en diferentes condiciones climáticas responden a perturbaciones climáticas similares de manera diferente (SAGREDO & LOWELL 2012). En los Andes Tropicales esta diferencia en la magnitud de la respuesta es muy visible debido a que el clima está influenciado por el Atlántico, el Pacífico y la circulación occidental de diversa magnitud. Esta magnitud de la influencia de los patrones de circulación varía desde los trópicos interiores hacia los trópicos exteriores y los resultados en diferentes patrones de precipitación y temperatura. En un clima de calentamiento, una temperatura alta en el invierno puede acelerar la ablación glaciar (BONNANO *et al.* 2013). KASER & OSMASTON (2002) propusieron que si la ubicación de un glaciar está por encima de la media anual de la isoterma 0°C, sería altamente sensible a la variabilidad de la precipitación e insensible a la variación de la temperatura. Los glaciares que se acumulan en el verano son más sensibles a la variación de temperatura que aquellos con acumulación de invierno (FUJITA 2008). Los trópicos interiores son cada vez más nublados y húmedos que los trópicos exteriores que son cada vez más secos (VUILLE *et al.* 2008b).La altitud es también otro factor importante que controla el retroceso de los glaciares en respuesta al cambio climático (CHEVALLIER *et al.* 2011). VEETTIL *et al.* (2016b) observaron que la tasa de aumento/disminución de la línea de nieve de los glaciares situados a mayor altitud fluctúa menos en comparación con los que están a altitudes menores. Esto es porque los glaciares de altitudes mayores en esta región son alimentados por fuertes nevadas (la precipitación máxima se produce en el verano en bajas latitudes tropicales), mientras que la fusión rápida del hielo se produce en los de partes más bajas (CHEVALLIER *et al.* 2011) y la temperatura del aire también es menor en glaciares de altitudes mayores en comparación con los glaciares de baja altitud. En un entorno de calentamiento, cuando no se produce ningún cambio en la precipitación, los glaciares más pequeños en las altitudes más bajas desaparecen más rápido debido a la disminución de la relación de acumulación/ablación (CHEVALLIER *et al.* 2011). La gran variabilidad en el balance de masa encontrado en el caso del glaciar Zongo en Bolivia demuestra la vulnerabilidad de los glaciares de baja altitud durante las últimas décadas (SORUCO *et al.* 2009).

A pesar de que las líneas de nieve han variado con las fases frías y calientes, no sería suficiente considerar solamente el ENOS para describir la variación en los cambios del clima de los glaciares en los Andes Tropicales. El Pacífico ecuatorial fue anormalmente cálido durante el periodo entre 1976 y 1997; esta fase de calentamiento se encuentra en la ODP positiva durante 1979-1997 (Fig. 10). Una tendencia de enfriamiento se encuentra durante 1948-1975, 1998-2002 y 2008-2014. La ocurrencia de El Niño se hizo más frecuente durante las últimas décadas y los regímenes fríos y calientes de la ODP se han desplomado. Los SST promedio se han incrementado drásticamente después de 1975, lo cual explicaría por qué el glaciar se ha retraído durante 1976-1997. Si se producen episodios de La Niña durante el régimen de calentamiento de la ODP, la ganancia en el balance de masas sería menor en comparación con la que se produce durante el frío régimen de ODP. Esto explicaría por qué los acontecimientos más fríos de La Niña no pudieron aportar un balance de masa positivo durante el reciente régimen cálido de la ODP (1975-1997). Durante los años de El Niño, los glaciares de Bolivia (Cordillera Oriental, trópicos húmedos del sur) fueron sometidos a un fuerte balance de masa negativa (WAGNON *et al.* 2001). Esto hizo que los glaciares tropicales están en una condición desequilibrio en el marco del clima actual (FAVIER *et al.* 2004). A pesar que existen limitaciones al utilizar el fenómeno ENOS para predecir los impactos futuros, este ayuda a calcular las anomalías climáticas (MASON & GODDARD 2001) con el uso de tablas de contingencia. Con excepción de pequeñas diferencias, los cuatro glaciares estudiados mostraron una respuesta similar a la ocurrencia de ENOS. Otra característica notable es que glaciares diferentes mostraron diferentes tasas de respuesta, especialmente en diferentes zonas climáticas. Esto significa que se necesita calibración específica para cada caso (JOMELLI *et al*. 2009). Existe un retraso de tres meses en las anomalías tropicales media zonal en relación con ecuatoriales SST del Pacífico oriental (KUMAR & HOERLING 2003; VUILLE *et al*. 2000b). En este estudio, se consideró un retraso de tres meses basado en VUILLE *et al*. (2000b), que explican claramente el retraso en los cambios en los patrones de precipitación y temperatura y se espera un retraso mayor hacia la Cordillera Oriental (Bolivia y Perú). Diferencias significativas en el comportamiento de los glaciares con un clima forzado y una estacionalidad de la precipitación (y por lo tanto de fusión) fue encontrada en los trópicos interior y exterior sobre la base de un estudio en el glaciar Zongo en Bolivia y el glaciar Antisana en Ecuador (FAVIER *et al*. 2004). De los resultados obtenidos de este estudio, se observa que las fluctuaciones de los glaciares fueron dominados por fases ENOS (junto con régimen de calentamiento de la ODP) y que el glaciar está desapareciendo gradualmente en correlación con el aumento de las variaciones recientes de SST. Este aumento reciente de SST en el Pacífico es debido al régimen de calentamiento de ODP y aumento de la frecuencia de ocurrencia de El Niño.

Estudios recientes sugieren mayores tasas de calentamiento a mayores altitudes (BRADLEY *et al*. 2006; RANGWALA 2013; PEPIN *et al*. 2015), el llamado calentamiento dependiente de la elevación (EDW). Los factores que explican este fenómeno son: mecanismos de nieve - albedo, la nubosidad, la mayor humedad, los aerosoles atmosféricos, o una combinación de todos estos juntos (PEPIN *et al*. 2015). Como se describe en la introducción, a pesar de que existe un hiato reciente en el calentamiento global (VUILLE *et al*. 2015) en las zonas costeras (2008 - 2015), las zonas altas los Andes Tropicales se siguen calentando y esto es evidente por el aumento de línea de nieve durante este período. A pesar de que la mayor humedad es una de las razones (RANGWALA 2013) detrás de EDW, se necesita más investigación para entender si otros factores también contribuyeron a las variaciones observadas en la línea de nieve en la Cordillera Blanca (trópicos exteriores húmedos del norte) y las cordilleras orientales de Bolivia y Perú (trópicos exteriores húmedos del sur). Además, PEPIN *et al*. (2015) propusieron la hipótesis de que las variabilidades interanuales a decadales en la circulación a gran escala, tales como ENOS y ODP, pueden influir en la variación regional en EDW, sobre todo en las zonas tropicales montañosas (DIAZ *et al*. 2014).

La modulación de la radiación solar es una fuerza externa que influye en las fluctuaciones de los glaciares (JOMELLI *et al*. 2009; VEETTIL *et al*. 2016b), la cual no es considerada en muchas investigaciones. Un aumento de la radiación solar entrante debido a la disminución de la nubosidad puede influir en la sensibilidad de los glaciares hacia un cambio en uno o muchos de los parámetros, tales como la temperatura del aire que influyen en la estabilidad de balance de masa (RABATEL *et al*. 2013). El máximo glacial en Perú y Bolivia y el mínimo solar de Maunder (1645-1715) se produjeron simultáneamente. Durante un poco significativo mínimo solar de Dalton (1783-1830), el retroceso de los glaciares en los trópicos se ha ralentizado (RABATEL *et al*. 2013). Sin embargo, muchos glaciares en los Alpes se han retirado o no avanzaron durante el Maunder mínimo (LUTERBACHER *et al*. 2001) y de hecho, el avance de glaciares a principios del siglo XIX en Europa fue más extenso que durante el mínimo de Maunder (ZASDANI 2007). Otros factores externos que influyen en las interacciones con el clima glaciar en los Andes Tropicales incluyen actividades volcánicas y la presencia de materiales particulados en la superficie de hielo. Actividades volcánicas son frecuentes en la región andina y pueden influir en el comportamiento de los glaciares debido directamente a glacio-química y a la reducción del albedo debido a los materiales particulados. Influencia indirecta debido a los aerosoles volcánicos en la estratosfera se discute más adelante en este documento.

Cambios glaciares en los trópicos interiores

Una pérdida de casi el 60% de la superficie del glaciar durante 1975 - 2015 es significativa en el caso de Cotopaxi en Ecuador. Para este, el valor de la ALN, el cual está cerca de LE del año, es difícil de obtener debido a la ausencia de una estación específica de precipitación, y esto junto con las discrepancias en los datos de teledetección, podría causar errores adicionales en el cálculo de la ALN. El volcán Antisana, que está justo por encima de la cuenca del Amazonas y cerca del Cotopaxi, perdió 33% de su superficie glaciar entre 1979 y 2007 (RABATEL *et al*. 2013). El volcán Chimborazo también perdió 57% de su zona de hielo (27,7 km2 a 11,8 km2 durante 1962-1997) (CÁCERES 2010).

La variación de la temperatura en los Andes ecuatorianos es dependiente de las SST Pacífico ecuatorial central (Niño - 3 y Niño - 3.4). Las pequeñas diferencias en las anomalías de precipitación (VEETTIL *et al*. 2014) encontradas en el caso de Chimborazo (Cordillera Occidental), Antisana y Cotopaxi (Cordillera Oriental) podrían explicar cómo el efecto ENOS varía espacialmente en las diferentes cordilleras. El balance de masa positiva observado por FRANCOU *et al*. (2004) en el Antisana - 15 durante la fase 1999-2000 de La Niña puede explicarse por la presencia de un régimen frío de ODP que ha prevaleció durante 1998-2002 (1997-1998 fue un periodo fuerte de El Niño). En este estudio, se observó que había una pérdida total de la superficie del hielo en los trópicos interiores pero la pérdida no fue continua, particularmente cuando se produjeron episodios de La Niña durante la fase fría del ODP. FRANCOU *et al*. (2003) describieron cómo el balance de masas está vinculado con fases ENOS, junto con los ciclos estacionales, donde la mayoría de las variaciones del balance de masa ocurren durante los meses de verano. Sin embargo, se encontró que la anomalía de las fases caliente/fría del ENOS afecta a la ablación del glaciar después de cuatro meses de su ocurrencia (CADIER *et al*. 2007). Entre 1984-1986 y 1999-2001 las fases de La Niña se impusieron y el glaciar mostró un lento retroceso. Largas fases cálidas a partir de 1986 y 2002 (El Niño + tibia ODP) han agotado la zona de hielo que fue adquirida durante el período de La Niña anterior y los años de El Niño se asociaron con precipitación por debajo del promedio en la mayoría de los casos (VUILLE *et al*. 2000a). La tasa de precipitación y la temperatura del aire en la superficie (y por lo tanto ENOS también) son los factores de control del equilibrio de masa. Se encontraron excepciones debidas probablemente a factores climáticos locales. En algunos casos en Ecuador y el norte de Perú, las precipitaciones pueden estar por encima del valor normal durante El Niño y por debajo de lo normal durante La Niña (MASON & GODDARD 2001). Los resultados obtenidos fueron similares a los reportados por WAGNON *et al*. (2001), basado en el estudio de glaciar Zongo en Bolivia para el período 1997-1998 (El Niño). Se encontró que el balance de masa mensual y la precipitación están más correlacionados durante los períodos de La Niña que durante períodos de El Niño (FRANCOU *et al*. 2004).

Trópicos exteriores húmedos del norte

Los cálculos existentes de la ALN y de la LE en la Cordillera Blanca incluyen Artesonraju entre 2000 y 2010 (RABATEL *et al*. 2012) y para el período 2011 - 2012 (DÁVILA 2013), y el glaciar Shallap (GURGISER *et al*. 2013) para el período 2006 - 2008. Muchos otros estudios calculan los cambios en el área (SILVERIO & JAQUET 2005; RACOVITEANU *et al*. 2008; VUILLE *et al*. 2008b; BURNS & NOLIN 2014) en el Cordillera Blanca, pero los resultados en las variaciones de la línea de nieve son comparables con los resultados de estos estudios. Se conoció que algunos glaciares en la Cordillera del Vilcanota en Perú pudieron haber perdido 32% de la superficie glaciar durante 1962 - 2006 (SALZMANN *et al*. 2012) y una pérdida de 35% en la parte sur de la Cordillera Blanca entre 1962 y 1999 (MARK & SELTZER 2005). Alrededor del 25% de la zona de los glaciares se perdió entre 1987 (643,5 km2) y 2010 (482 km2), y los glaciares situados hacia el sur perdieron un mayor porcentaje de su superficie en comparación con sus contrapartes del norte en la Cordillera Blanca (BURNS & NOLIN 2014). RACOVITEANU *et al*. (2008) calcularon una disminución del 22,4% en la zona del glaciar entre 1970 (723,3 km2) (AMES *et al*. 1989), y 2003 (569,6 km2) en la misma área de estudio. La mayor pérdida de áreas glaciares en el sur (BURNS & NOLIN 2014) puede deber a su menor altitud o debido a la proximidad de la ZCIT de enero como sugieren los autores. Recientemente, ALARCÓN *et al*. (2015) también confirmaron el rápido retroceso de los glaciares en el extremo sur de la Cordillera Blanca (área de estudio) y observaron una pérdida de superficie de alrededor de 58% entre 1975 y 2010. Las pequeñas discrepancias en la cuantificación de los cambios en la superficie pudieron surgir de las diferencias en la resolución espacial de las imágenes utilizadas. Se tiene información de que glaciares en Bolivia se han retirado rápidamente entre 1975 y 1983, y de nuevo entre 1997 y 2006 (SORUCO 2008), por lo cual este estudio confirma esta conclusión. Se observó el balance de masa durante períodos de estabilidad entre 1956 - 1975 (régimen frío de la ODP) y 1992 - 1996 (inesperado). Tendencias similares fueron observadas en el caso de Antisana - 15 (FRANCOU *et al*. 2000, 2003) y Cotopaxi (VEETTIL *et al*. 2014) en los Andes ecuatorianos. Se encontró que una rápida retirada ocurrió en el caso del Nevado Cololo, cuando se compara con los glaciares en Ecuador.

En esta región (trópicos exteriores húmedos del norte), la temperatura del aire estacionalmente es uniforme (cerca de 0°C) y la humedad media es de alrededor del 71%. SCHAUWECKER *et al*. (2014) observaron un incremento del calentamiento de 0,31°C/década entre 1969 y 1998, mientras el calentamiento se desaceleró a 0,13°C/década de 1983 y 2012 en la Cordillera Blanca, y recientemente, VEETTIL *et al*. (2016b) también observaron condiciones similares. Una pequeña disminución en la precipitación se observa cerca del Nevado Tuco, en particular durante JJA, de las últimas seis décadas, así como un ligero aumento en la precipitación DEF entre 2000 y 2010 (VEETTIL *et al*. 2016b). La disminución observada en la precipitación, sin embargo, no es suficiente para causar un mayor retroceso de los glaciares en esta región. Un aumento en la precipitación DEF junto con un aumento en la temperatura del aire puede contribuir a una mayor retracción de los glaciares debido a que las precipitaciones dominan en lugar de nieve en condiciones de calentamiento, que a su vez provocan una disminución de albedo de la superficie. Sin embargo, cabe señalar que las lluvias ocurren rara vez en elevaciones más altas de los trópicos exteriores, en particular cerca de las zonas subtropicales. MARK & SELTZER (2005) también observaron un fuerte calentamiento a lo largo de la Cordillera Blanca entre 1962 y 1999. Si el calentamiento provoca un aumento de la altitud de congelación, esto puede conducir a un mayor punto de fusión debido a que hay lluvia en lugar de nieve (BRADLEY *et al*. 2006). Las causas del aumento excepcional de la humedad desde la década de 2000, según lo observado por

VEETTIL *et al*. (2016b) no se entienden bien porque hubo una pequeña anomalía de temperatura negativa entre 2000 y 2008, y la humedad es mayor que las variaciones esperadas de la temperatura (entre 2000 y 2008). El aumento de la humedad, en lugar de los cambios de temperatura, se cree que es la causa del aumento de línea de nieve excepcional en esta región, debido al cambio de pérdida de masa debida a sublimación por pérdida de fusión. El aumento de la humedad antes de finales de 1990sin embargo, era consistente con el calentamiento observado (VUILLE *et al*. 2003; MARK & SELTZER 2005). En comparación con otro estudio (VEETTIL *et al*. 2016a) sobre las variaciones de línea de nieve en los trópicos exteriores secos, se observa que los glaciares de la Cordillera Blanca están bajo una mayor amenaza de retroceso rápido, sobre todo después de 2010. Esta observación encaja en el aumento excepcional de humedad en lugar de aumento de la temperatura o disminución de la precipitación. Por otra parte, los glaciares en los trópicos exteriores son de respuesta relativamente lenta a los cambios de temperatura en general en comparación con los glaciares en los trópicos interiores (FAVIER *et al*. 2004).

El cambio prolongado en régimen negativo de la ODP (1947-1976) a su régimen positivo a finales de 1970 podría haber contribuido al calentamiento observado desde 1979 hasta 1998, en los Andes Tropicales (GARREAUD *et al*. 2009). Además, los años más fuertes de El Niño en 1997-1998 junto con muchos otros episodios de El Niño entre 1991 y 1995 contribuyeron a estos aumentos de las condiciones de calentamiento. Una fuerte correlación entre ENOS y el balance de masa/energía de la superficie (SMB/SEB) ya está establecido en la Cordillera Blanca, basado en un estudio del glaciar Shallap (MAUSSION *et al*. 2015) y esta correlación fue cuantitativamente mayor en comparación con los estudios anteriores (VUILLE *et al*. 2008b). La posición geográfica de la Cordillera Blanca (con respecto a la región Niño - 3.4) sin duda influye en este aumento de la correlación entre los cambios en los glaciares y ENOS. Además, al ser relativamente lejos de la ZCIT de Enero y siendo menos influenciado por los patrones de circulación del Amazonas y del Atlántico (SAGREDO & LOWELL 2012; SAGREDO *et al*. 2014), la influencia del Pacífico determina la mayor parte del clima en la Cordillera Blanca. Es conocido en los Andes tropicales que cuando ODP se produce en fase con ENOS, la influencia de ENOS en glaciares es más fuerte y más visible que cuando ENOS está en fase opuesta a la ODP o se produce durante periodos neutros de ODP (VEETTIL *et al.* 2014, 2015, 2016a). Se observó un débil correlación entre la variabilidad anual de precipitación y ODP en la Cordillera Blanca durante los años recientes (SCHAUWECKER *et al*. 2014) y esto podría influir en cambio del balance de masa también (VEETTIL *et al.*2015, 2016a). Sin embargo, una correlación fuerte ocurrió entre la temperatura del aire y ODP antes de alrededor de 1995 (SCHAUWECKER *et al*. 2014) después del cambio de régimen de ODP en 1976. Las señales de este fenómeno, a pesar de que no es uniforme, son también visibles en las variaciones de línea de nieve en la Cordillera Blanca.

Trópicos exteriores húmedos del sur

Las variaciones de la ALN del Nevado Cololo mostraron una tendencia al alza con altas fluctuaciones. VEETTIL *et al*. (2015) observaron que, sin embargo, la línea de nieve del nevado Sajama en los trópicos secos exteriores fluctúa de manera acentuada con la ocurrencia del ENOS y la ODP en comparación con el Nevado Cololo. La elevación de la línea de nieve durante los episodios de El Niño durante 1991-1995 debe ser evaluado cuidadosamente. Hubo un aumento rápido en la línea de nieve durante el período 1991-1992 (Fig. 9), mientras que la línea de nieve no varió mucho en el período 1992-1995 a pesar de que los eventos de El Niño fueron más fuertes y prolongados, siendo esta observación válida para todos los demás sitios de estudio, excepto para el Cotopaxi debido a la falta de datos. Una de las posibles causas de esta anomalía puede ser el efecto de enfriamiento de los aerosoles volcánicos en la estratosfera debido a la erupción del monte Pinatubo el 15 de junio de 1991 (RABATEL *et al*. 2013; VEETTIL *et al*. 2015, 2016a).

RAPER & BRAITHWAITE (2006) y RUPPER & ROE (2008) sugieren que los glaciares en el clima húmedo son más sensibles a la elevación de temperatura que en un clima seco. La fuente principal de humedad para la precipitación en el Altiplano proviene de las tierras bajas orientales de los Andes, la cual es altamente dependiente de las anomalías de SST tropicales en el Atlántico (VUILLE *et al*. 2000). Los eventos de El Niño inducen déficit de precipitación en los trópicos exteriores, donde la precipitación líquida (lluvia) está casi ausente en las zonas altas, que promueve la fusión de los glaciares. Aparte de las variaciones de altitud, la tasa de pérdida de masa glaciar en los Andes Tropicales es altamente dependiente de características geométricas tales como la exposición. Se encontró que glaciares orientados hacia el este y al sur tuvieron menor retroceso en comparación con aquellos orientados hacia el norte y al oeste en la Cordillera Real, cerca del Nevado Cololo (SORUCO *et al*. 2009).

La variabilidad interanual del balance de masa en el sitio de estudio depende altamente de la variabilidad interanual de la precipitación durante el verano (FAVIER *et al*. 2004) y la variabilidad de la precipitación interanual es altamente dependiente de las anomalías de la circulación atmosférica durante las fases extremas de la Oscilación del Sur (VUILLE *et al*. 2000). Si el glaciar está cubierto de nieve, que tiene alto albedo, la ablación rápida puede prevenirse durante el tiempo de precipitación débil (SORUCO *et al*. 2009). La precipitación y la temperatura del aire son factores determinantes en los cambios de la línea de nieve en los trópicos exteriores.

La precipitación y temperatura en el sitio de estudio fueron altamente influenciadas por ENOS debido a la reducción en la advección de aire húmedo del continente provocada por vientos del oeste de nivel medio (VUILLE 1999; VUILLE *et al*. 2000a; GARREAUD & ACEITUNO 2001). La circulación de Walker es uno de los rasgos definitorios de clima tropical, una fuerte circulación de Walker implica una condición La Niña y uno más débil implica una condición de El Niño. Debido a baja resolución espacial de los conjuntos de datos sobre precipitación, la correlación entre el ENOS y la variabilidad de la precipitación no se estableció en este estudio. La correlación entre el ENOS y la precipitación durante DEF en el Altiplano ha sido referida como alta (GARREAUD *et al*. 2009). Durante los eventos de El Niño, en el caso del Nevado Cololo, una mayor precipitación mensual se produce debido al transporte aéreo de la húmeda de la cuenca del Amazonas. Se sabe que prevalecen condiciones de frío y seca a lo largo de la costa del Pacífico, y condiciones relativamente cálidas y húmedas prevalecen en el continente (GARREAUD 2009). Registros de los núcleos de hielo de la capa de hielo en Quelccaya - Perú sugieren que un aumento de la temperatura se ha producido durante el siglo XX (THOMPSON *et al*. 1984), a pesar de los cambios en la precipitación no son tan visibles (VUILLE *et al.* 2008a).

Trópicos exteriores secos

Debido a su condición seca y fría, la variación en línea de nieve en la Cordillera Occidental (trópicos exteriores secos) no es tan rápida como el caso del Nevado Cololo (trópicos exteriores húmedos). Además, el lado oeste del altiplano es más sensible a ENOS de la región oriental (VUILLE *et al*. 2000b). La disminución general en la zona glaciar del Nevado Coropuna durante 1986 - 2014 (26,92%) calculado por VEETTIL *et al*. (2015) fue muy similar a la que ocurrió durante 1962 - 2000 (26%) calculado por RACOVITEANU *et al*. (2007). Una recesión significativa de los glaciares en los Andes del Perú comenzó a mediados del siglo XIX (KASER 1999). KASER *et al*. (1996) estudiaron la relación entre los cambios en la línea de equilibrio de los glaciares andinos y las fluctuaciones en el clima. Sin embargo, la presencia de la capa de nieve en exceso en el frente glaciar fue siempre un obstáculo para delinear los límites del glaciar y línea de equilibrio utilizando datos de teledetección en esta región y este problema fue posteriormente superado por el uso de línea de nieve como indicador del clima (ARNAUD *et al*. 2001). La presencia de estaciones secas y húmedas distintas hace que sea posible calcular con mayor precisión línea de nieve anual con la utilización de imágenes de satélite en los trópicos exteriores secos en comparación con los de trópicos interiores. Las imágenes de satélite tomadas durante el final de la temporada seca (Mayo - Septiembre) ya han demostrado ser excelentes en el cálculo de la línea de nieve anual que a su vez puede ser utilizado como sustituto de la LE con base en los estudios de caso sobre glaciar Zongo de Bolivia y glaciar Artesonraju de Perú, por RABATEL *et al*. (2012).

Sobre la base de los modelos de circulación general, MINVIELLE & GARREAUD (2011) calcularon una disminución significativa en la circulación del Este sobre el Altiplano y que puede causar una fuerte disminución de la precipitación en los trópicos centrales andinos hacia el final del siglo XXI. La variabilidad interdecadal de los Andes está asociada con cambios a largo plazo en los patrones de circulación del Pacífico, mientras que la variabilidad decadal está asociada con los cambios en los patrones de circulación de la cuenca Amazónica (ESPINOZA-VILLAR *et al*. 2009). A pesar de que muchos de los modelos existentes, tales como CMIP3, la tendencia inconsistente en la precipitación regional limita severamente la comprensión del cambios climáticos a lo largo del altiplano (MINVIELLE & GARREAUD 2011). Sin embargo, ya es entendido que el SST del Pacífico ha aumentado desde la década de 1970, debido al llamado cambio climático Pacífico y este podría ser uno de los agentes causantes del retroceso acelerado observado en los glaciares (RABATEL *et al*. 2013).

HERREROS *et al*. (2009) mencionaron que no hubo cambio alguno en la cantidad de precipitación en el área de estudio durante el evento de El Niño en 1997-1998, pero el aumento de la línea de nieve fue mayor al de los resultados obtenidos en esta investigación. Durante las temporadas fuertes de El Niño, los valores de temperatura fueron más altos en la región Coropuna (VEETTIL *et. al*. 2016a).

Vientos del Oeste (condiciones secas) y vientos del Este (condiciones de humedad) son las anomalías que no están centradas sobre la parte central del altiplano y la ubicación de estas anomalías en los vientos son importantes para determinar el patrón espacial de las anomalías de precipitación en los Andes centrales. Un estudio en cuencas de drenaje en Perú muestra que la región costera está teniendo mayor variabilidad en las precipitaciones y por tanto mayor variabilidad en la escorrentía en escalas de tiempo estacionales e interanuales (LAVADO-CASIMIRO *et al*. 2012). Está claro que sí existe un aumento en la escorrentía durante la temporada de baja precipitación, debido únicamente a la ablación acelerada de los glaciares. Incluso, aunque no hay documentación pertinente sobre la humedad, existen registros en los Andes. VUILLE *et al*. (2008b)encontraron un aumento moderado de humedad relativa entre 1950 y 1995 en la parte occidental de Bolivia, basados en datos de la estación. Existe una alta correlación entre la pérdida de masa glaciar y la temperatura del aire en las latitudes medias y altas latitudes glaciares (BRAITHWAITE 1981) y se explica cómo los eventos de El Niño fueron seguidos por una elevación de la línea de nieve anual.

El aumento excepcional de la línea de nieve anual entre 1997 - 1998, 2004 - 2005 y 2010 y la disminución de la superficie del glaciar (VEETTIL *et al*. 2016a) puede ser bien explicado por la presencia de eventos fuertes de El Niño ocurridos durante el régimen de calentamiento de la ODP. A partir de los resultados, se observó que tanto la superficie de los glaciares como de las nieves no variaron tanto como se esperaba por la influencia combinada del El Niño y la fase positiva de la ODP que se produjo durante 1991 - 1995; esto es similar a los resultados observados en el Nevado Cololo en Bolivia. ARNAUD *et al*. (2001) reportaron que en el 2001 ocurrió una influencia de ENOS en la Nevado Sajama en Bolivia, donde también está situado el Nevado Coropuna.

CONCLUSIÓN

Los glaciares tropicales andinos están retrocediendo a un ritmo alarmante desde finales de 1970. Sin embargo, la tasa de retroceso no fue uniforme a lo largo de los Andes Tropicales. Se observó una mayor tasa de retiro en los trópicos exteriores húmedos del sur y los trópicos interiores, en tanto que los glaciares de los trópicos secos exteriores presentaron tasas de retiro inferiores. A pesar de que se necesitan más investigaciones, el presente estudio indica que los glaciares de las cordilleras orientales de Perú y Bolivia están retrocediendo a mayor velocidad. Esto puede causar una disminución en el agua dulce de origen glaciar en algunas ciudades, como en La Paz, Bolivia. Como consecuencia inmediata de un mayor retroceso de los glaciares, no se puede pasar por alto la probabilidad de ocurrencia de una catástrofe relacionada a glaciares como el GLOF en algunas cadenas montañosas como en la cordillera de Apolobamba.

A partir de la comparación de ENOS con las variaciones en el máximo anual de las líneas de nieve (aproximadamente igual a la línea de equilibrio anual), se observaron diferentes tasas de subida/caída de la línea de nieve durante los períodos de El Niño y La Niña. Sin embargo, se observó la influencia de ENOS en los glaciares modulados por las fases fría y caliente de la ODP. Cuando El Niño se produce durante el calentamiento ODP, la línea de nieve puede aumentar hasta varios metros. Una vez más, se observó la tendencia de las variaciones de la línea de nieve a ser diferente en los trópicos interiores y los trópicos exteriores. Una rápida disminución de la extensión de los glaciares se produjo durante 1975-1997, en paralelo a la fase de calentamiento de la ODP.

No se observaron signos del reciente hiato en el calentamiento global en el comportamiento de los glaciares. Sin embargo, como se ha mencionado en estudios recientes, la región subtropical fue testigo del hiato en el calentamiento global y la tasa de retroceso de los glaciares recientes (2008-2015) fue menor en los trópicos exteriores secos que están cerca de las zonas subtropicales.

Los datos de velocidad del viento y de sublimación no fueron utilizados en este estudio. Sobre una base a largo plazo, cuando estos se combinan con los registros de núcleos de hielo, se puede obtener una imagen sin ambigüedades. Actualmente se está pasando por un período de pruebas combinadas de El Niño y ODP cálida. La comprensión de los cambios de los glaciares en las actuales condiciones climáticas es de gran relevancia en este contexto.

AGRADECIMIENTOS

Esta investigación fue financiada por La Fundação de Amparo à Pesquisas do Estado do Rio Grande do Sul (FAPERGS).

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CAPITULO IX

Conclusões e recomendações

1. CONCLUSÕES

Os registros de geleiras tropicais andinas não só nos ajudam a compreender o clima no passado, mas também proporcionam uma visão do que está acontecendo no clima atual, bem como as tendências futuras. As alterações da linha de equilíbrio são amplamente utilizadas nos Andes tropicais para compreender as interações modernas entre o clima e as geleiras. A linha de neve máxima, derivada das imagens de satélite durante a estação seca, serve para calcular o valor aproximado da linha de equilíbrio e é usada extensivamente para compreender as variações no balanço de massa das geleiras andinas tropicais. Observou-se uma maior taxa de recuo das geleiras nos trópicos internos e trópicos externos úmidos do sul, enquanto nos trópicos externos as taxas de recuo apresentado foram inferiores.

Os trópicos internos e os trópicos externos mantêm as condições de temperatura homogêneas ao longo do ano com uma pequena sazonalidade (na temperatura do ar) nos trópicos externos. A acumulação é limitada à estação de precipitação nos trópicos externos e ocorre em todo o ano nos trópicos internos. Existem gradientes na resposta das geleiras às alterações climáticas que ocorrem dos trópicos internos para os trópicos externos nos Andes (norte a sul). As variações observadas na altitude da linha de neve nesta pesquisa não são homogêneas ao longo dos Andes tropicais (Veettil et al., 2014, 2016a, 2016b, 2016c. 2016d). As geleiras tropicais dos trópicos internos (Venezuela, Colômbia e Equador), especialmente as situadas perto da Zona de Convergência Intertropicais (ZCIT) de janeiro, são mais vulneráveis a aumentos na temperatura, sendo menos sensíveis a variações na precipitação. Em contraste, as geleiras dos trópicos externos respondem à variabilidade de precipitação muito rapidamente em comparação com a variação de temperatura, em especial quando se deslocam para as regiões subtropicais. As observações sobre os gradientes na resposta das geleiras às alterações climáticas nos Andes tropicais neste estudo estão em consonância com o novo quadro de a classificação (Sagredo e Lowell, 2012; Sagredo et al., 2014) das geleiras (como trópicos internos, trópicos externos úmidos do norte, trópicos externos úmidos do sul e trópicos externos secos).

Através da comparação de ocorrências de ENOS e as variações de altitude das linhas de neve, foram observadas diferentes taxas de subida ou descida nas linhas de neve durante os períodos de El Niño e La Niña (Veettil et al., 2014, 2016a, 2016b, 2016c. 2016d). No entanto, observou-se que a influencia de ENOS em geleiras foi modulada por fases quentes e frias ODP (Veettil et al., 2014, 2016a). A rápida diminuição na extensão glacial ocorreu durante 1975-1997, paralelamente à fase de aquecimento ODP. Quando El Nino ocorre durante ODP quente, a altitude da linha de neve pode subir vários metros. Mais uma vez, observou-se a tendência de diferentes variações na altitude da linha de neve nos trópicos internos e trópicos externos. Observa-se que as geleiras acima, e perto da ZCIT (em janeiro) (trópicos internos e trópicos externos úmidos do sul), estão recuando mais rápido (Veettil et al., 2016d). Já as geleiras nos trópicos externos úmidos do norte e trópicos externos secos mostraram recuo relativamente mais lento (Veettil et al., 2016b, 2016c). Possivelmente isso pode acontecer devido à ocorrência simultânea das fases frias do El Niño – Oscilação Sul (ENOS) e da Oscilação Decenal do Pacífico (ODP). As anomalias observadas nas variáveis meteorológicas seguem padrões ODP, e as variações anuais de linha de neve seguem os padrões de eventos de El Niño, particularmente quando na fase com ODP quente. No entanto, uma forte correlação entre as variações de linha de neve e ENOS (e ODP) não está estabelecida.

A comparação entre as geleiras do Equador e as da Colômbia e Venezuela (Ceballos et al., 2006; Morris et al., 2006; Braun e Bezada, 2013) mostram que no Equador (Veettil et al., 2014) o recuo das geleiras é menor devido, provavelmente, à grande altitude das geleiras equatorianas. Em geral, observa-se um rápido desaparecimento das geleiras menores dos trópicos internos e trópicos externos úmidos do sul situadas em baixas altitudes, diferentemente do que ocorre com as demais geleiras nos Andes tropicais (Veettil et al., 2014, 2016a, 2016d). Esta pesquisa é um avanço na glaciologia Andina tropical, que considera as variações das geleiras em uma escala continental.

Quando analisadas as mudanças na área das geleiras nos locais de estudo (exceto no Equador), em uma escala de décadas, observa-se que a área coberta de gelo situado nas encostas orientais está recuando mais rapidamente do que aquelas geleiras nas encostas ocidentais. Se essa tendência continuar, poderia ter implicações hidrológicas nesses rios nos lados orientais dos Andes, incluindo alguns rios da bacia amazônica brasileira, que estão alimentadas por geleiras. Esta observação, sobre a propriedade direcional do recuo das geleiras, deve ser analisada com cuidado, pois o recuo reforçado de gelo nos lados leste e nordeste podem depender de outros fatores, tais como: a quantidade de precipitação líquida (chuva) devido à influência da Bacia Amazônica, e das declividades mais elevadas no lado leste. Esta hipótese deve ser testada em toda a cordilheira oriental para verificar se todas as geleiras mostram esta propriedade direcional, ou se esta propriedade está restrita apenas para as geleiras analisadas neste estudo.

A dependência de balanço de massa nas características de sublimação também aumenta a partir dos trópicos internos para os trópicos externos (Veettil et al., 2016d). As condições de aquecimento com maior umidade tendem a aumentar a perda de massa devido ao derretimento e não a sublimação (Wagnon et al., 1999). O aumento de umidade observado nos trópicos externos e subtrópicos podem alterar as geleiras que estão dominadas pela sublimação, fazendo com que futuramente sejam dominadas por derretimento. De fato, muitos estudos (Bradley et al., 2006; Liu et al., 2009; Rangwala et al., 2009; Rangwala, 2013; Pepin et al., 2015) já demonstraram que o aumento da umidade em ambientes de grande altitude é uma das causas do aquecimento nestas áreas (aquecimento dependente da elevação - ADE).

Não há sinais de recente hiato no aquecimento no comportamento das geleiras. No entanto, tal como mencionado em alguns estudos recentes (Schauwecker et al., 2014; Vuille et al., 2015), a região subtropical testemunhou o hiato e a taxa de recuo das geleiras recentes (2008-2015) foi menor nos trópicos externos secos que estão perto de regiões subtropicais.

2. RECOMENDAÇÕES

Apesar do recente hiato no aquecimento global (Vuille et al., 2015), tem-se observado um aumento contínuo do aquecimento nas regiões de montanha com grande elevação nos Andes (Veettil et al., 2016d). A inclusão da elevação como um parâmetro ao interpretar as interações entre as geleiras e o clima ao longo do Andes tropicais é importante neste contexto.

Não se sabe ainda por que as geleiras próximas a Bacia Amazônica, onde a maioria da precipitação continental ocorre, estão recuando mais rápido do que as geleiras mais próximas da costa do Pacífico. Um recuo mais elevado é observado na cordilheira oriental do Peru e da Bolívia em comparação com as geleiras na Cordilheira Branca e assim, são necessárias mais pesquisas para entender essas diferenças.

Investigações adicionais também são necessárias para entender a direção do recuo das geleiras nos Andes tropicais e a generalização dos resultados desta pesquisa. É importante também desenvolver um modelo para o recuo das geleiras e integrar com modelos hidrológicos, pois estas análises não foram consideradas neste projeto de pesquisa.

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ANEXO A

Alterações na área total de geleiras por década durante o período de 1975-2015

Tabela 1: Cotopaxi

Ano	A imagem usada	Referência	Superfície (km^2)
1976	Fotografias aéreas	Jordan *et al*. (2005)	19,24
1987	Landsat TM	Este estudo	13,65
1997	Fotografias aéreas	Jordan *et al*. (2005)	13,45
2001	Landsat ETM+	Este estudo	12,76
2014	Landsat OLI	Este estudo	11,37

Tabela 2: Nevados Tuco-Pastoruri

Ano	A imagem usada	Referência	Superfície (km^2)
1975	Landsat MSS	Este estudo	25.17
1987	Landsat TM	Este estudo	16.69
1998	Landsat TM	Este estudo	14.72
2005	Landsat TM	Este estudo	13.07
2015	Landsat OLI	Este estudo	11.05

Tabela 3: Nevado Coropuna

Ano	A imagem usada	Referência	Superfície (km^2)
1975	Landsat MSS	Este estudo	66.01
1986	Landsat TM	Este estudo	57.57
1996	Landsat TM	Este estudo	52.57
2006	Landsat TM	Este estudo	48.63
2015	Landsat OLI	Este estudo	43.04

Tabela 4: Nevado Cololo

Ano	A imagem usada	Referência	Superfície (km^2)
1975	Landsat MSS	Este estudo	43.69
1985	Landsat TM	Este estudo	38.71
1996	Landsat TM	Este estudo	31.22
2005	Landsat TM	Este estudo	24.97
2015	Landsat OLI	Este estudo	22.97

ANEXO B

Mudanças na área das geleiras em diferentes faixas de elevação

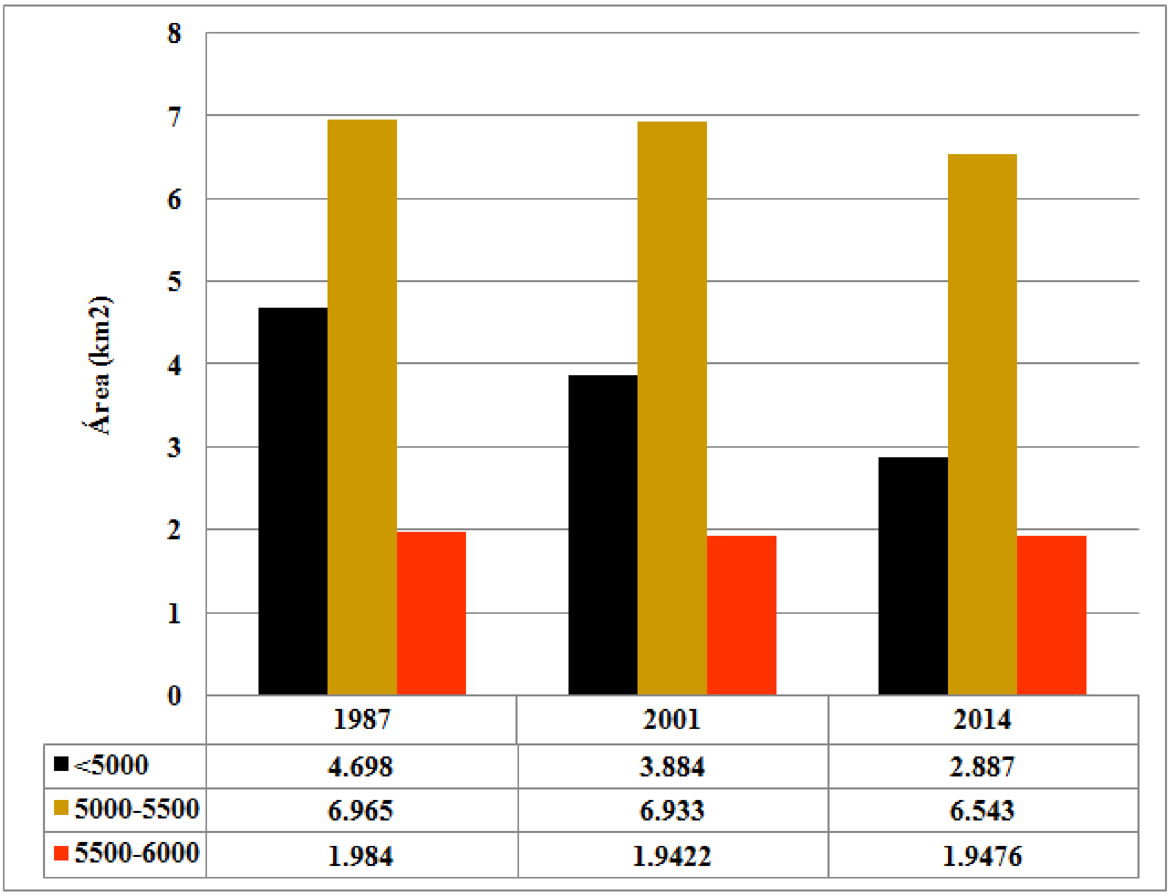
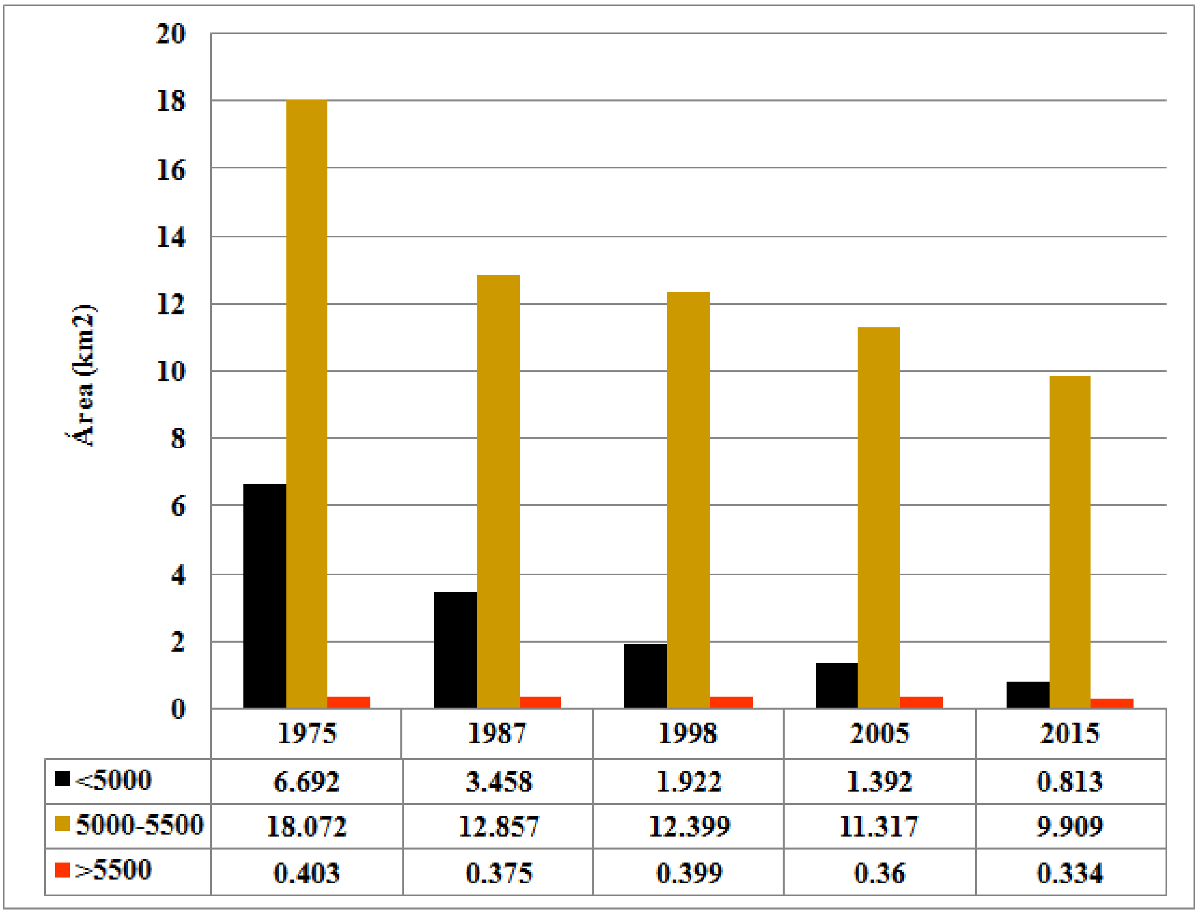
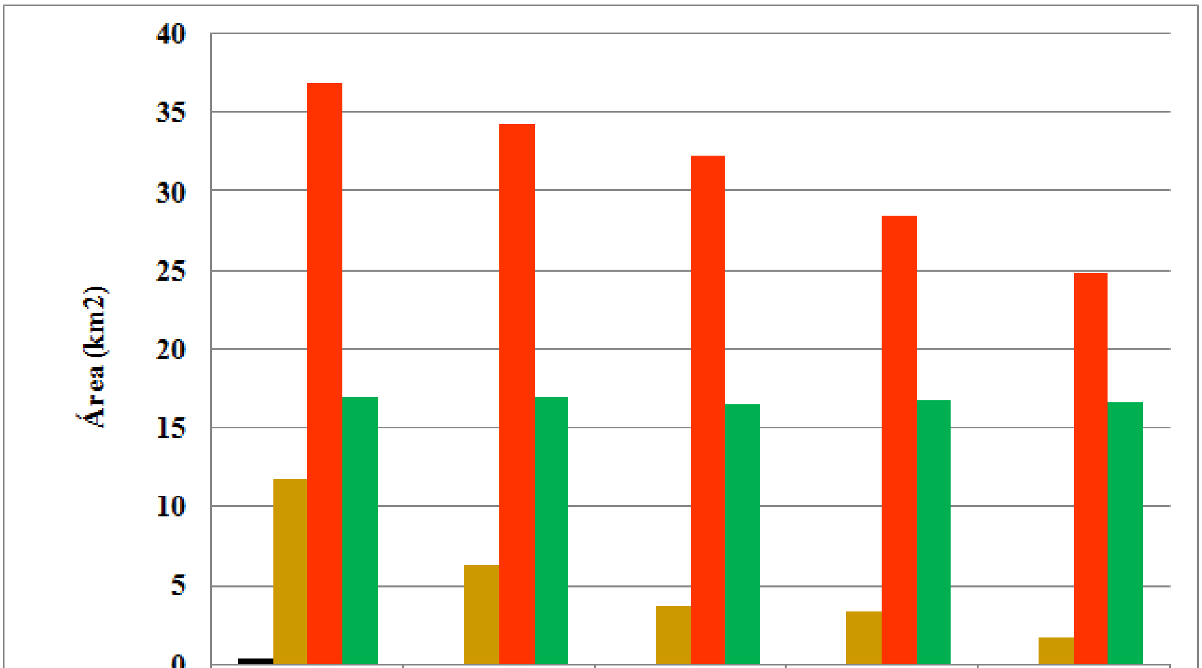


Figura 1: Mudanças na área de Cotopaxi em diferentes faixas de elevação



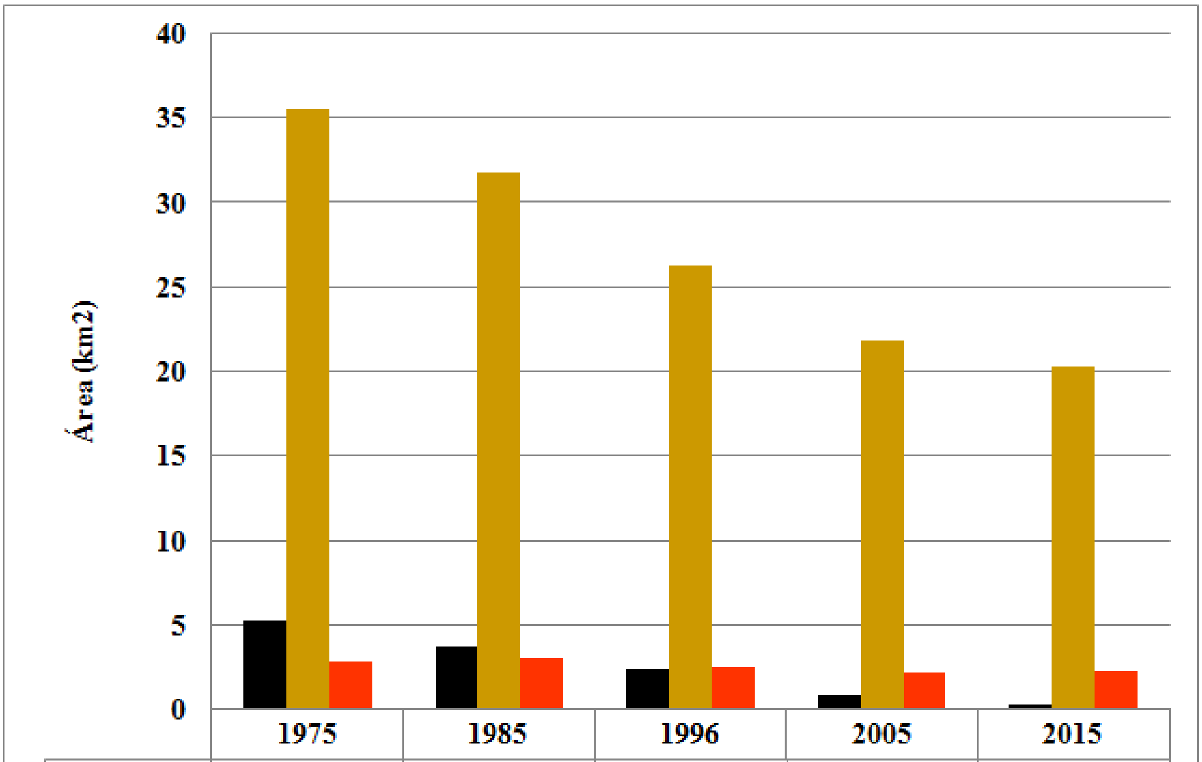
	1975	1987	1998	2005	2015
<5000	6.692	3.458	1.922	1.392	0.813
5000-5500	18.072	12.857	12.399	11.317	9.909
>5500	0.403	0.375	0.399	0.36	0.334

Figura 2: Mudanças na área de Nevados Tuco-Pastoruri em diferentes faixas de elevação



	1975	1986	1996	2006	2015
<5000	0.346	0	0	0	0
5000-5500	11.719	6.35	3.733	3.336	1.644
5500-6000	36.958	34.248	32.317	28.537	24.792
>6000	16.989	16.97	16.54	16.756	16.608

Figura 3: Mudanças na área de Nevado Coropuna em diferentes faixas de elevação



	1975	1985	1996	2005	2015
<5000	5.285	3.772	2.398	0.905	0.342
5000-5500	35.577	31.835	26.26	21.835	20.32
>5500	2.833	3.107	2.564	2.229	2.311

Figura 4: Mudanças na área de Nevado Cololo em diferentes faixas de elevação

ANEXO C

Mudanças na área das geleiras em diferentes aspectos

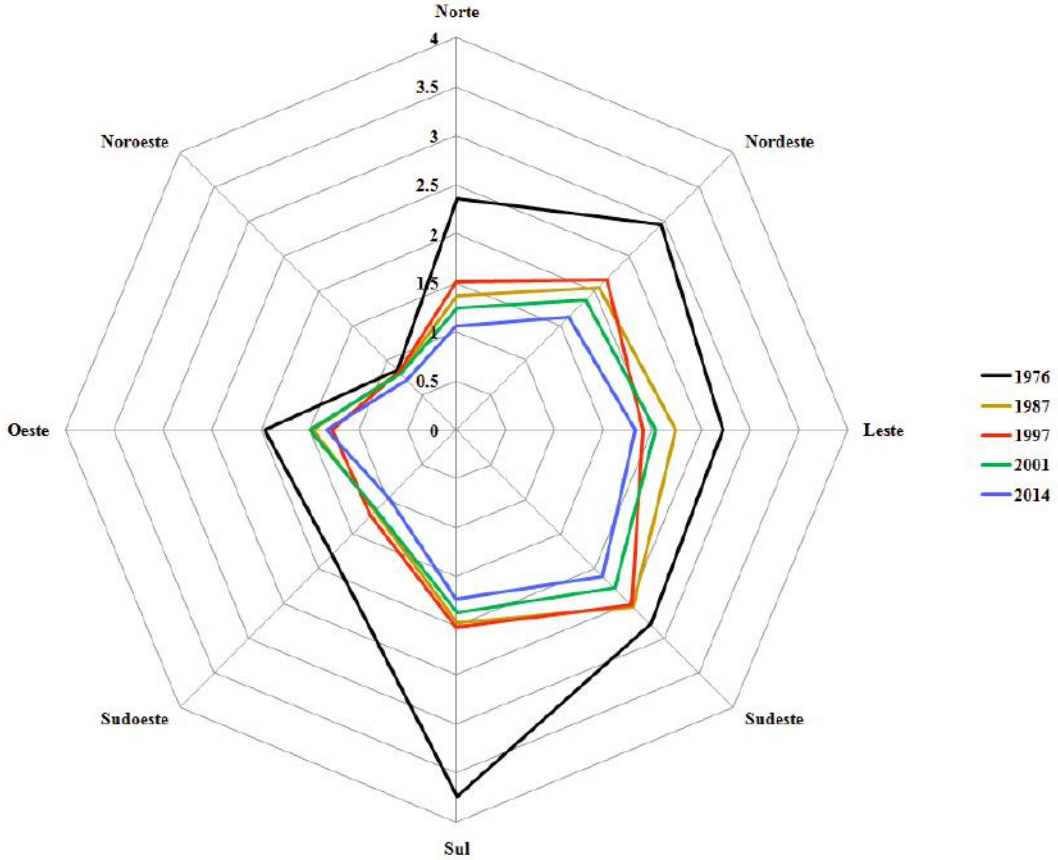


Figura 1: Mudanças na área de Cotopaxi em aspectos diferentes (km^2)

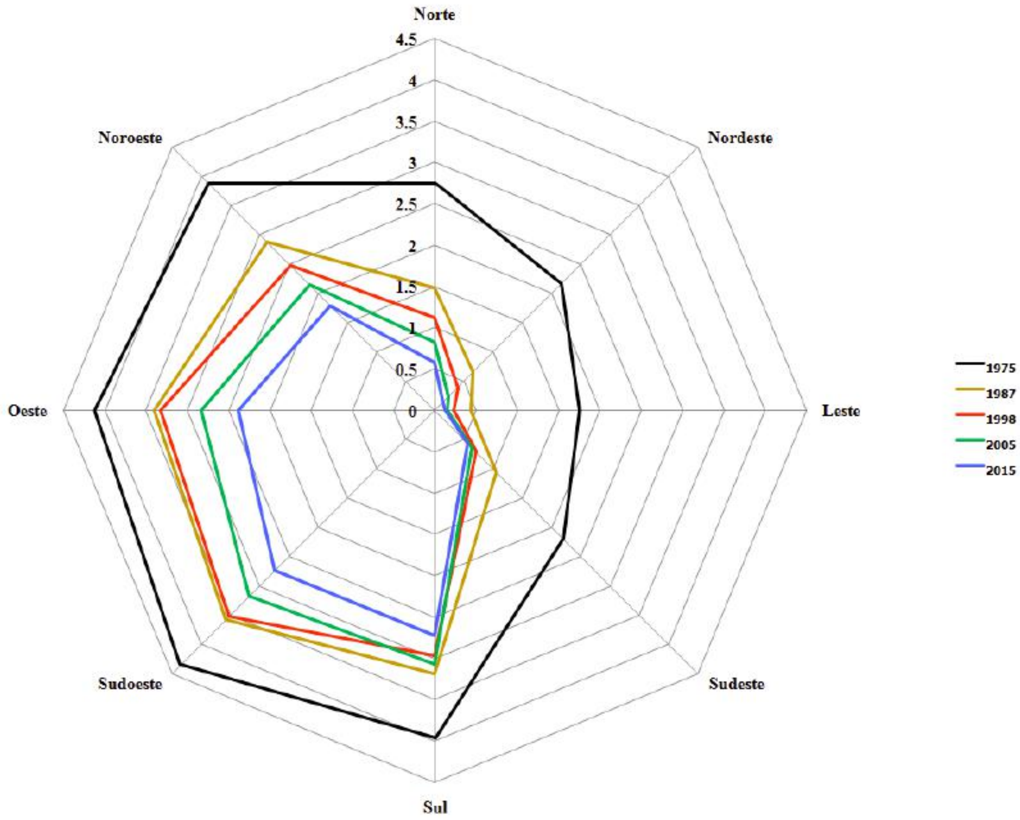


Figura 2: Mudanças na área de Nevados Tuco-Pastoruri em aspectos diferentes (km^2)

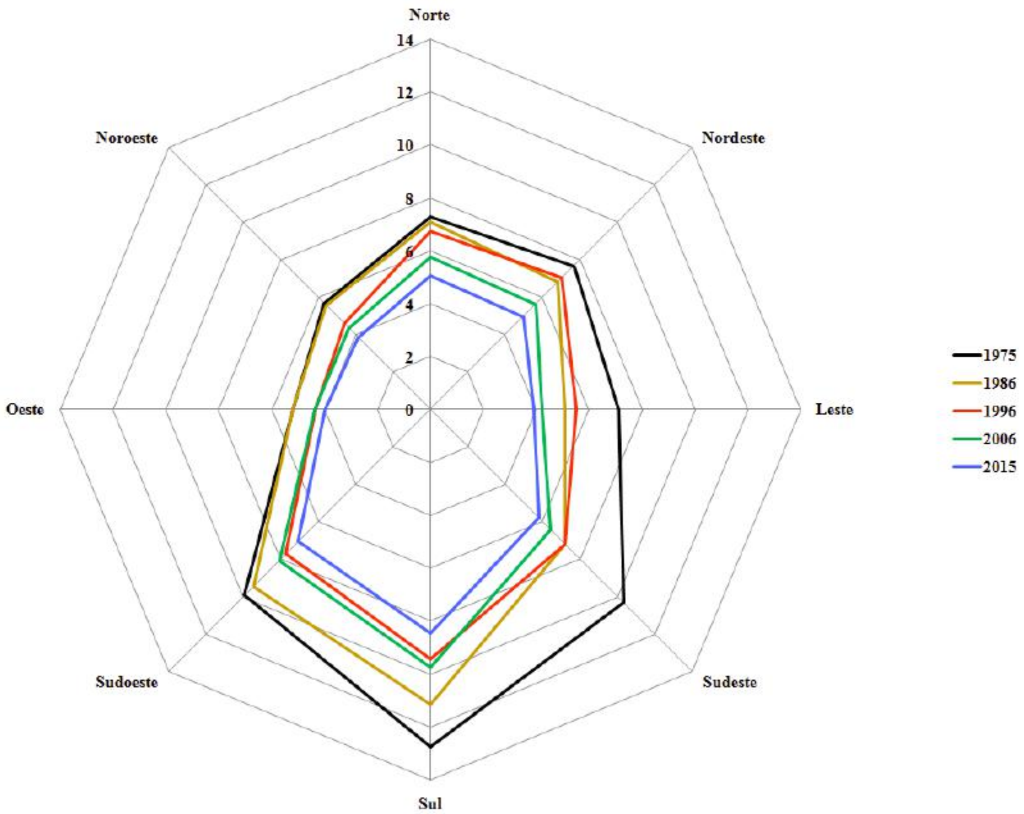


Figura 3: Mudanças na área de Nevado Coropuna em aspectos diferentes (km^2)

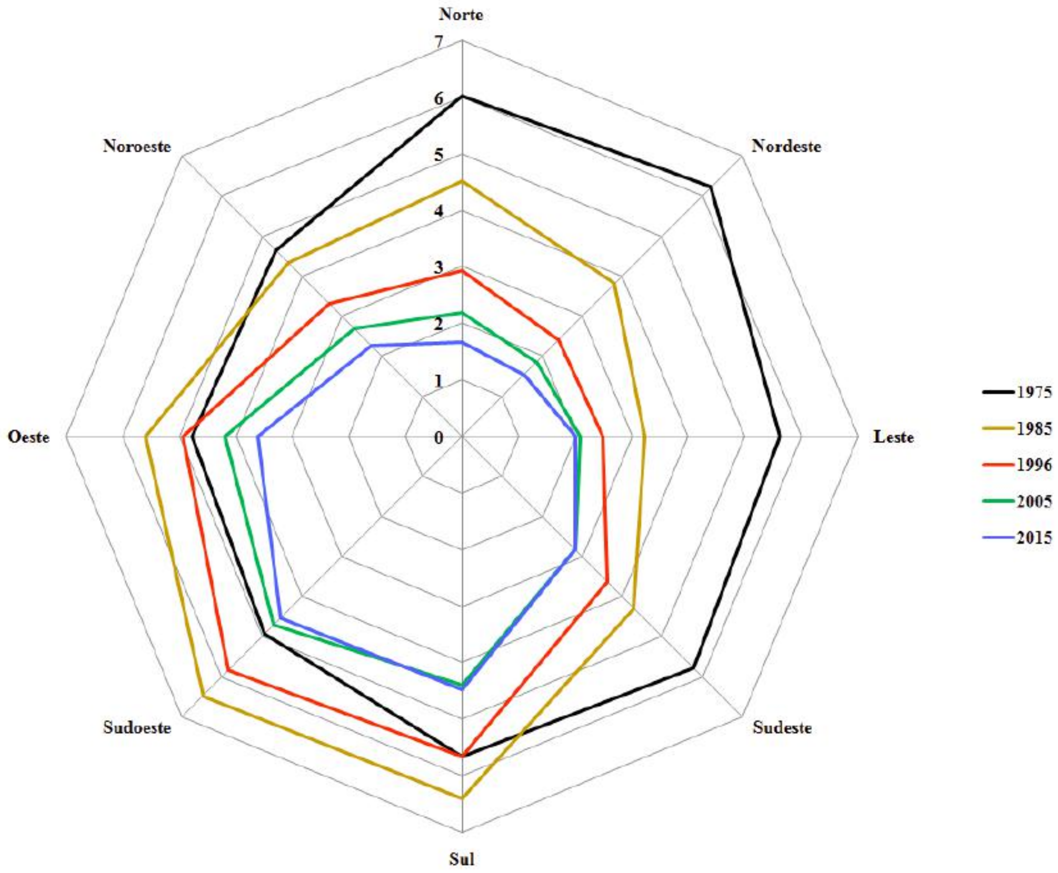


Figura 4: Mudanças na área de Nevado Cololo em aspectos diferentes (km^2)

ANEXO D

Linha de neve calculada na forma de tabelas (gráficos publicados em revistas)

Tabela 1: Cotopaxi

Ano	ALN-1	ALN-2	ALN-3
1979	4801	4684	4654
1987	4858	4744	4717
1988	4860	4738	4717
1989	4878	4772	4732
1991	4904	4769	4736
1996	4805	4749	4727
1997	4918	4786	4741
1999	4941	4787	4741
2000	4903	4777	4741
2001	4915	4777	4741
2004	4894	4786	4741
2005	4970	4811	4751
2006	4970	4849	4747
2007	4981	4878	4764
2008	4953	4874	4764
2009	4970	4867	4757
2010	4960	4851	4751
2012	4901	4867	4764
2013	4915	4867	4769

Tabela 2: Nevado Tuco-Pastoruri

Ano	ALN-1	ALN-2	ALN-3	ALN-4	ALN-5	ALN-6	ALN-7	ALN-8	ALN-9	ALN-10	ALN-11
1984	4931	4809	4936	4869	5020	4927	4893	4934	4892	4824	4961
1985					5029		4903	4939	4897		
1986	4931	4818	4940	4874	5029	4927	4900	4939	4897	4833	4971
1987	4935	4824	4940	4882	5033	4927	4910	4938	4897	4833	4978
1988	4945	4827	4943	4901	5039	4930	4911	4940	4901	4842	4979
1989	4945	4818	4940	4901	5036	4930	4911	4940	4901	4845	4985
1990	4945	4813	4940	4910	5036	4930	4911	4939	4901	4845	4979
1991	4950	4824	4943	4930	5040	4941	4934	4940	4901	4864	4987
1992	4952	4831	4952	4938	5044	4948	4942	4944	4906	4884	4987
1993	4953	4827	4950	4953	5044	4938	4942	4940	4908	4884	4983
1994	4953	4824	4951	4956	5044	4948	4942	4940	4906	4878	4987
1995	4957	4831	4951	4967	5045	4954	4942	4946	4906	4889	4988
1996	4955	4827	4951	4969	5045	4948	4953	4946	4902	4895	4987
1997	4961	4857	4952	4967	5045	4954	4953	4946	4917	4901	4985
1998	4972	4862	4953	4969	5059	4965	4989	4958	4917	4913	4993
1999	4978	4832	4951	4965	5051	4961	4989	4946	4910	4915	4992
2000	4972	4843	4944	4967	5051	4961	4989	4946	4910	4915	4992
2001	4986	4839	4951	4969	5059	4965	5011	4946	4906	4919	4985
2002	4986	4858	4951	4974	5062	4965	5011	4951	4917	4947	4987
2003	5000	4858	4953	4979	5066	4973	5029	4951	4937	4952	4989
2004	5012	4859	4958	4985	5073	4978	5029	4951	4946	4958	4992
2005	5020	4871	4960	4990	5080	4984	5029	4953	4971	4972	4998
2006	5032	4871	4968	4994	5080	4997	5044	4958	4971	4972	4999
2007	5042	4872	4975	4996	5080	5002	5048	4958	4981	4982	4999
2008	5042	4881	4974	4996	5077	4984	5047	4953	4985	4982	4997
2009	5042	4872	4969	4993	5077	4984	5047	4953	5000	4982	4997
2010	5038	4871	4962	4991	5077	4984	5047	4953	5000	4989	4997
2011	5038	4871	4961	4989	5077	4984	5047	4953	5000	4989	4997
2012	5053	4872	4961	4989	5080	4992	5051	4966	5000	4989	4997
2013	5053	4878	4961	4991	5080	4992	5051	4966	5000	4989	5005
2014	5060	4881	4967	4992	5080	5001	5055	4986	5000	5008	5013
2015	5067	4885	4967	4999	5084	5001	5055	4986	5000	5010	5017

Tabela 3: Nevado Coropuna

Ano	ALN-1	ALN-2	ALN-3	ALN-4	ALN-5
1986	5706	5435	5463	5509	5507
1987	5709	5441	5468	5525	5512
1988	5720	5442	5469	5530	5521
1989	5725	5448	5474	5534	5527
1990	5706	5441	5463	5521	5504
1991	5709	5451	5468	5525	5510
1992	5714	5455	5470	5534	5515
1993	5706	5450	5463	5530	5510
1994	5709	5445	5460	5522	5509
1995	5714	5460	5468	5535	5512
1996	5714	5457	5462	5536	5517
1997	5709	5455	5461	5528	5511
1998	5725	5480	5494	5538	5533
1999	5704	5449	5462	5520	5511
2000	5709	5480	5480	5529	5517
2001	5716	5494	5488	5536	5520
2002	5709	5480	5478	5528	5511
2003	5709	5507	5489	5543	5533
2004	5716	5510	5492	5555	5543
2005	5716	5513	5501	5557	5558
2006	5720	5518	5507	5561	5568
2007	5716	5518	5502	5559	5561
2008	5725	5527	5532	5588	5582
2009	5725	5537	5541	5594	5592
2010	5729	5564	5554	5619	5615
2011	5716	5550	5527	5565	5587
2013	5725	5564	5534	5569	5601
2014	5744	5576	5561	5578	5626

Tabela 4: Nevado Cololo

Ano	ALN-1	ALN-2	ALN-3
1985	5055	5004	4709
1986	5076	5016	4748
1987	5117	5043	4863
1988	5138	5086	4919
1989	5072	5028	4893
1990	5067	5003	4875
1991	5129	5081	4946
1992	5151	5110	5011
1993	5222	5146	5014
1994	5246	5162	5029
1995	5221	5146	4963
1996	5257	5162	5010
1997	5221	5135	4973
1998	5336	5220	5051
1999	5174	5090	4957
2000	5084	5016	4843
2001	5071	4989	4839
2002	5153	5044	4929
2003	5216	5135	4966
2004	5241	5135	4996
2005	5263	5150	5002
2006	5241	5137	4986
2007	5271	5151	4993
2008	5236	5122	4957
2009	5280	5183	4999
2010	5316	5226	5040
2011	5269	5170	4979

Tabela 5: Nevado Sajama

Ano	ALN-1	ALN-2
1984	5252	5121
1985	5364	5177
1986	5404	5202
1987	5566	5306
1988	5683	5244
1989	5535	5232
1990	5918	5460
1991	5518	5383
1992	5874	5453
1993	5898	5560
1994	5862	5522
1995	5894	5523
1996	5916	5535
1997	5920	5591
1998	6007	5619
1999	5644	5236
2000	5778	5499
2001	5481	5213
2002	5557	5262
2003	5854	5497
2004	5674	5499
2005	5572	5499
2006	5638	5532
2007	5940	5603
2008	5919	5560
2009	5862	5532
2010	5940	5582
2011	5894	5518

ANEXO E

Serie temporal (média anual) de precipitação e de temperatura (a 2 m do solo) nos locais de estudo baseados nos dados de NCEP-NCAR Reanalysis-2 e de CRU.

Precipitação

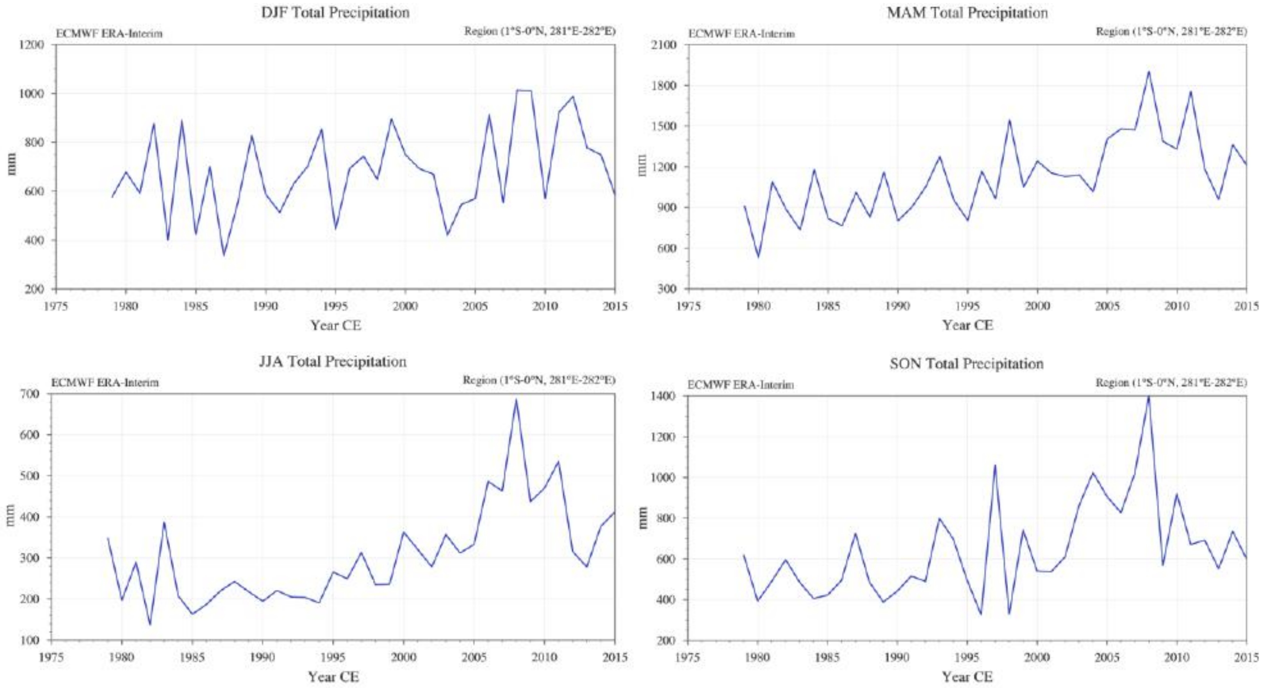


Figure 1: Cotopaxi

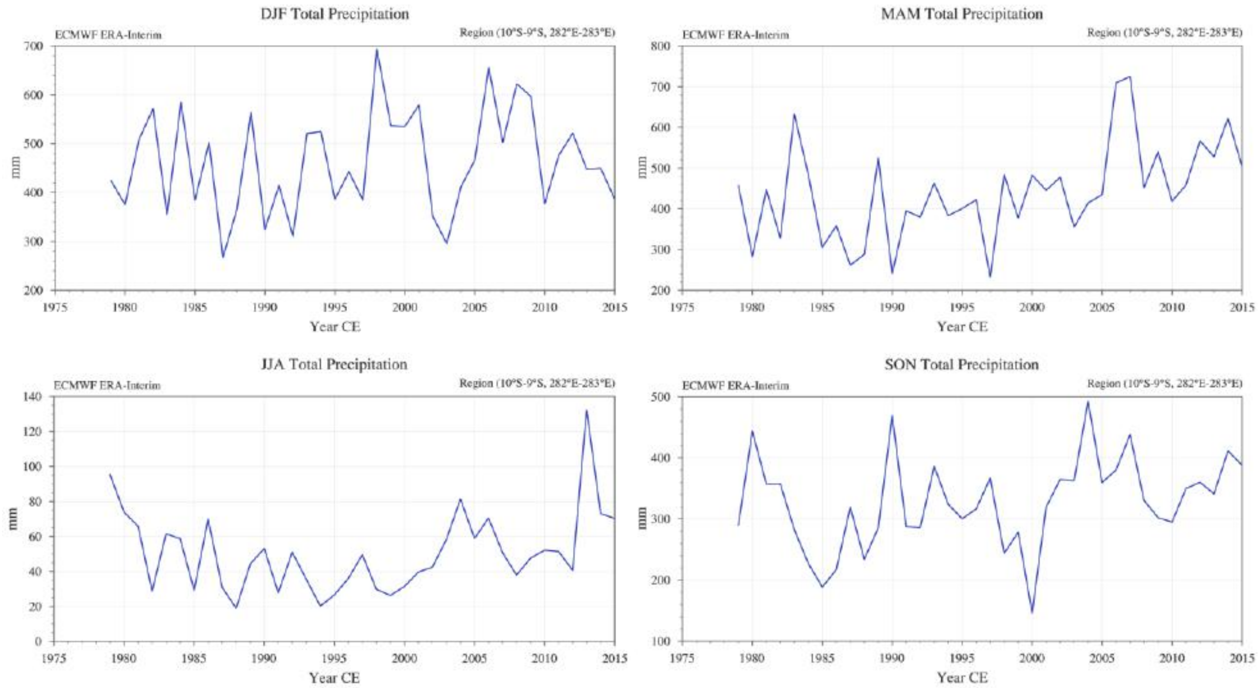


Figure 2: Nevado Tuco

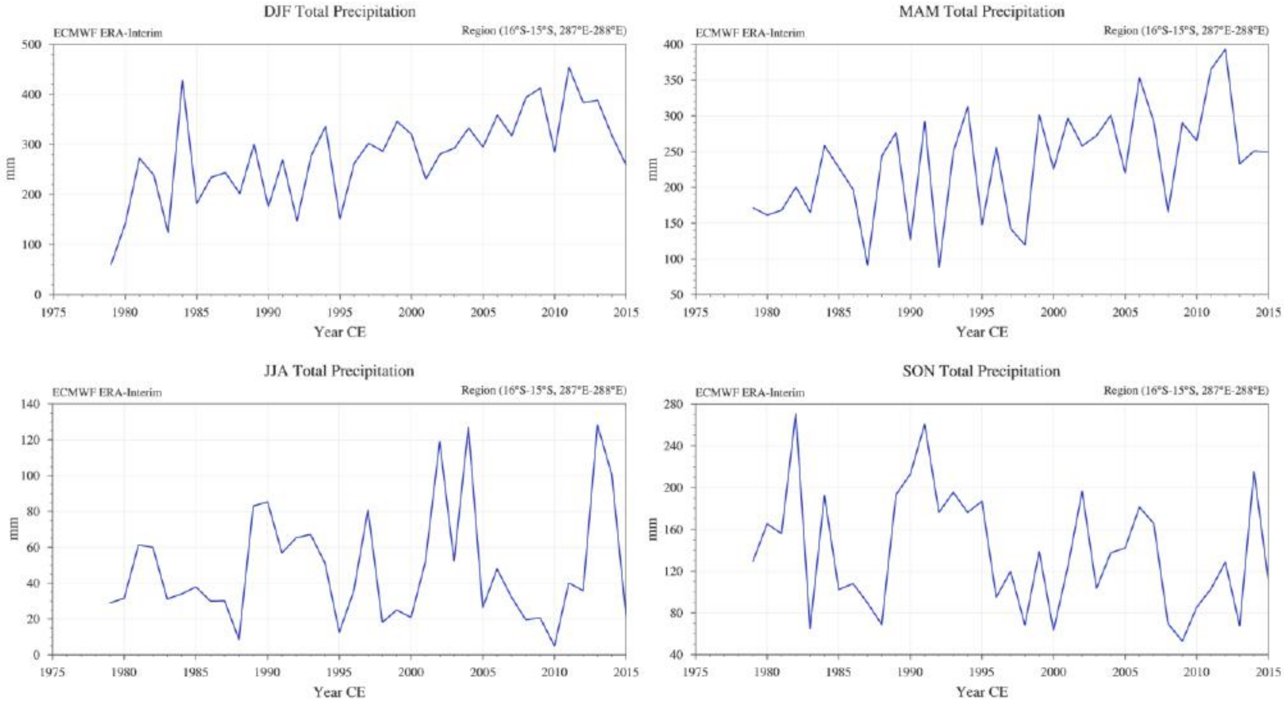


Figure 3: Coropuna

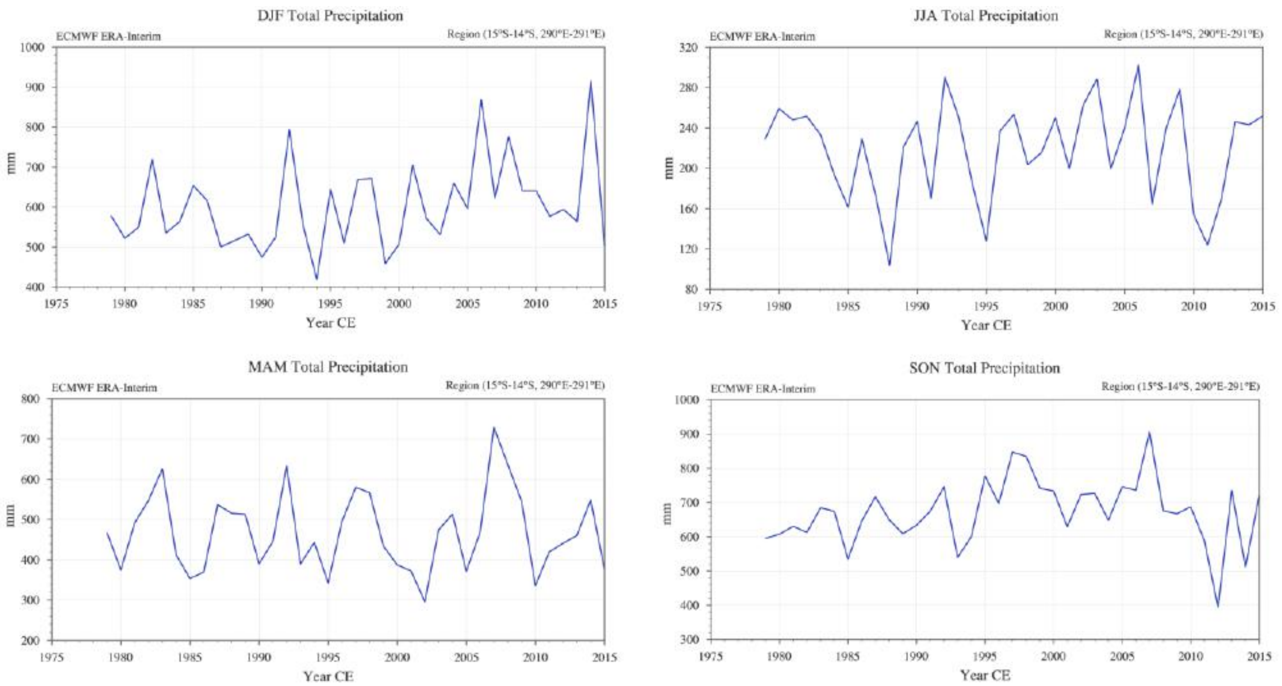


Figure 4: Cololo

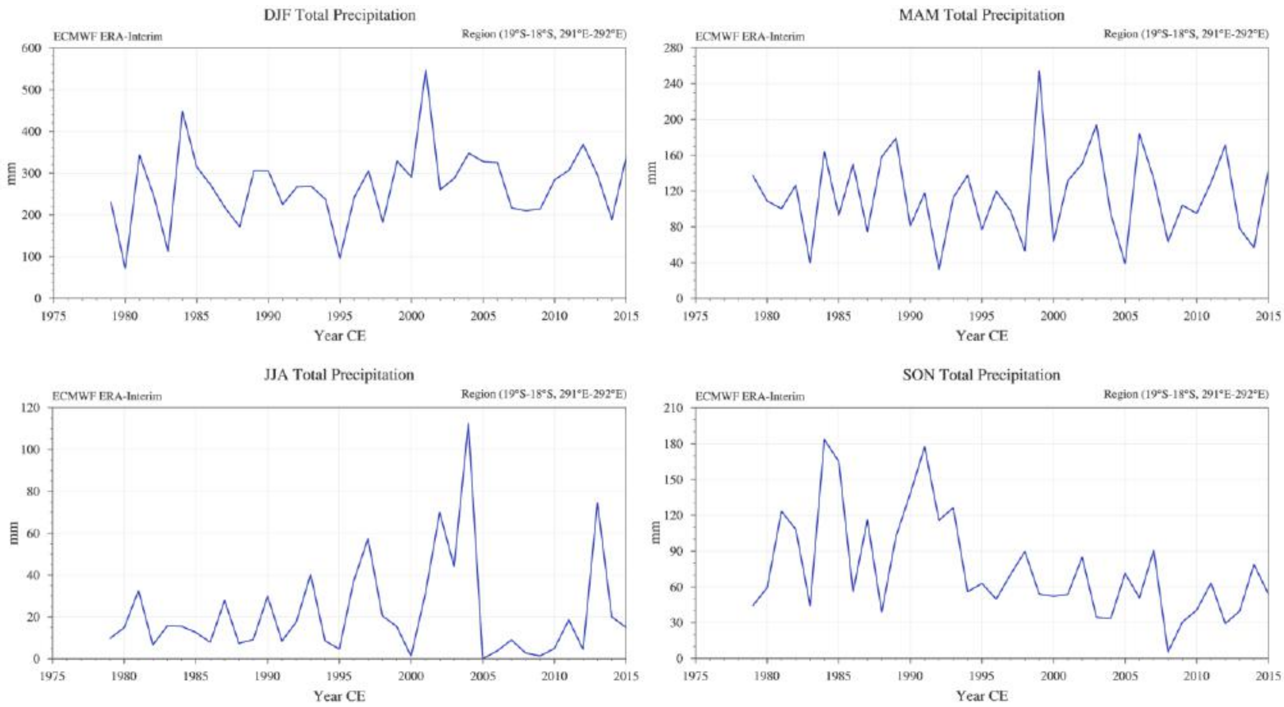


Figure 5: Sajama

Temperatura

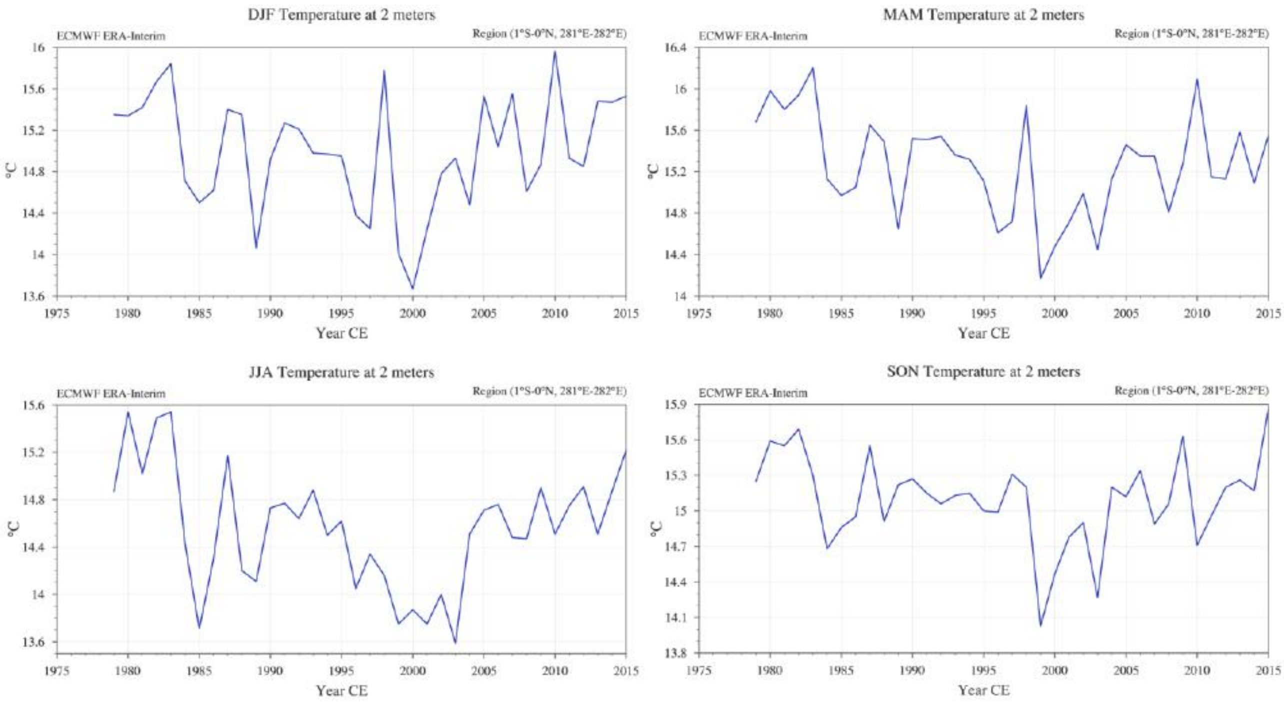


Figure 1: Cotopaxi

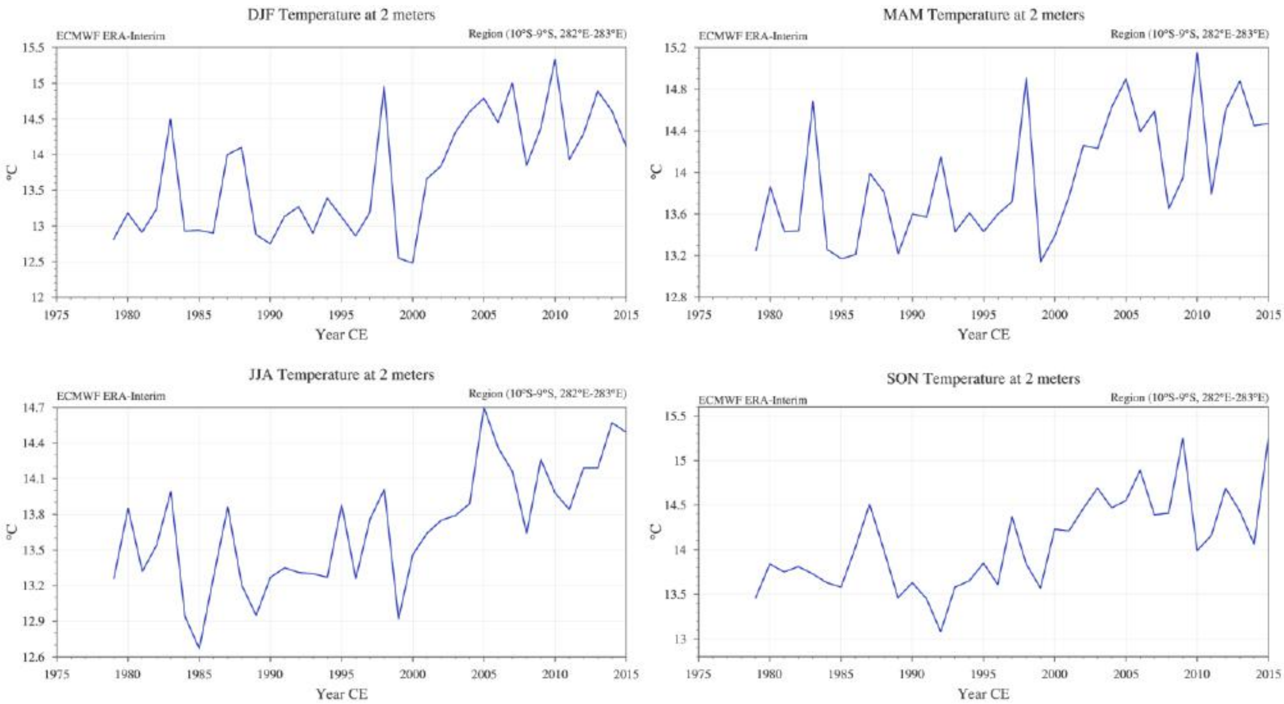


Figure 2: Tuco

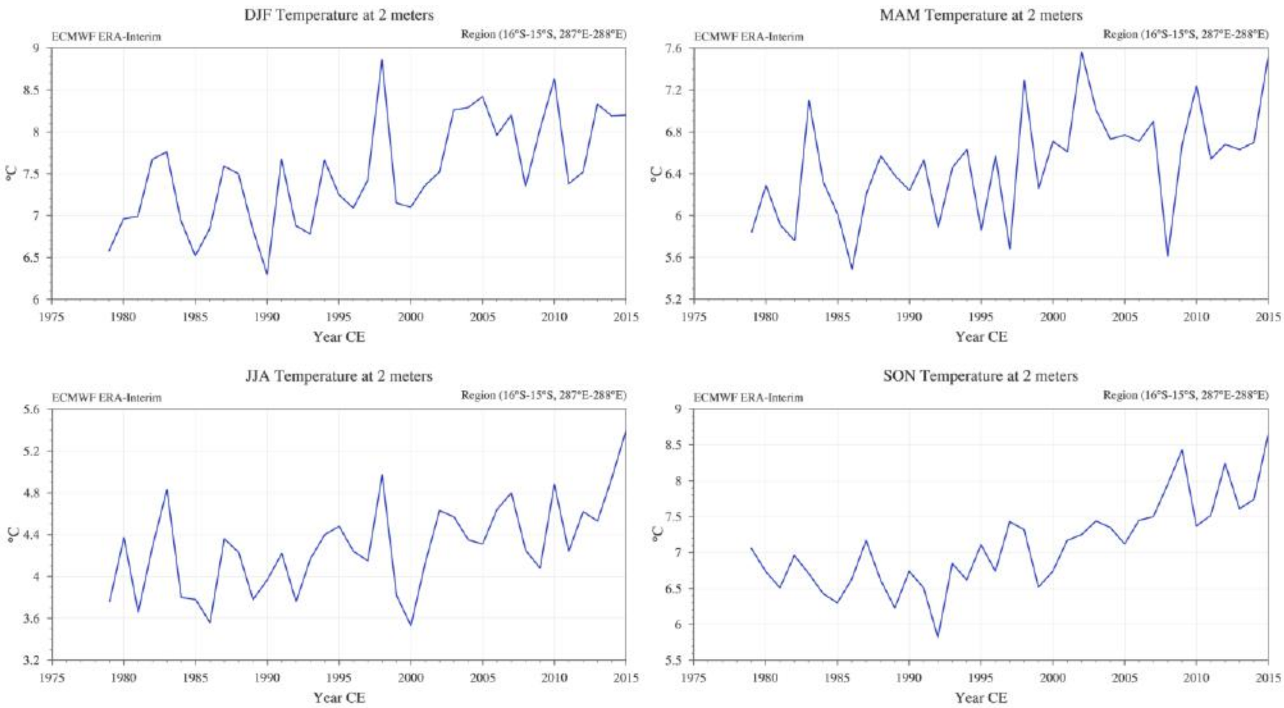


Figure 3: Coropuna

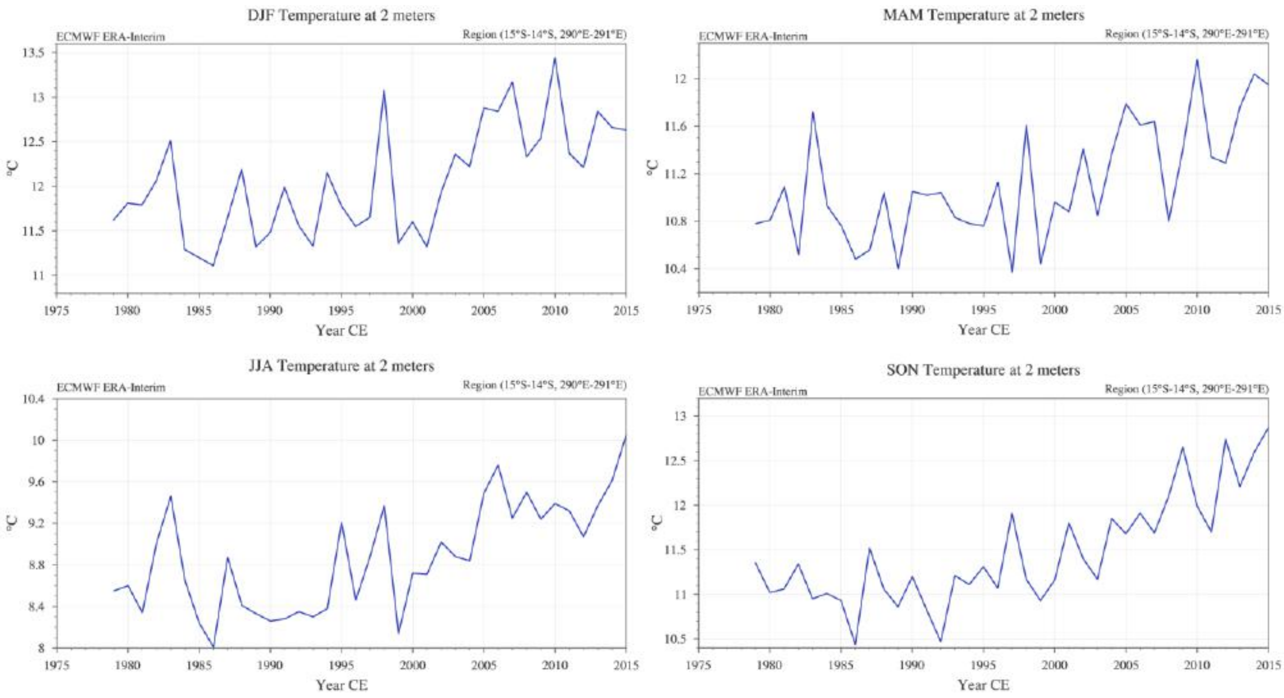


Figure 4: Cololo

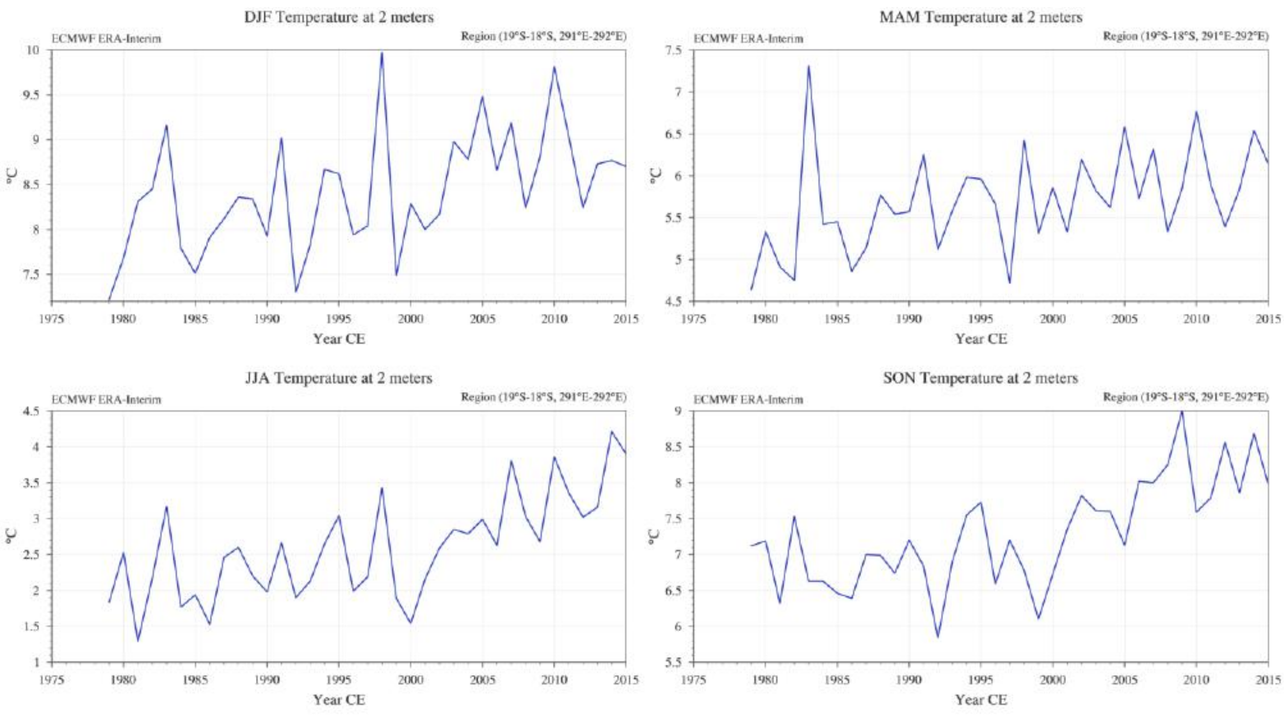


Figure 5: Sajama