Appendix 2-B. Bio-physical features of the tropical Pacific Ocean linked to changes in surface climate

Ocean currents

The circulation of the western and central Pacific Ocean (WCPO) is dominated by two broad westward currents, the North Equatorial Current (NEC) and the South Equatorial Current (SEC). These are separated by two eastward counter currents underneath the atmospheric inter-tropical and sub-tropical convergence zones (ITCZ and SPCZ) at latitudes 5°–10°N and 5°N–10°S. As the broad flows encounter islands and coasts, they form narrow and powerful currents that can transport heat, nutrients, particles and plankton over large distances.

Due to a strong relationship between winds and sea surface temperatures (SSTs), these ocean currents vary substantially with season and with the ENSO. In the central and eastern equatorial region, where the trade winds drive an upwelling of deep water and create the Pacific Equatorial Divergence Province (PEDP), the surface waters are relatively rich in nutrients.[1]

Most of the rest of the western tropical Pacific Ocean is nutrient-poor (oligotrophic) because warm surface water is piled up by the trade winds, pushing down the deep layers that are rich in nutrients. Wind-induced mixing, oceanic eddies, internal waves and island upwellings all act to reduce this natural stratification barrier supplying the upper layers of the ocean with the nutrients necessary for biological production. In the western equatorial Pacific, an immense pool of warm, nutrient-poor water moves back and forth along the equator according to the phase of the ENSO cycle. The eastern edge of the Warm Pool in the western Pacific defines the limit of the nutrient-rich PEDP.

The main currents in the tropical Pacific Ocean are driven by the easterly trade winds (Figure 1).[2] The primarily westward flow of these currents extends over the top few hundred metres of the ocean and reaches the western boundary of the basin. In the first few tens of metres, an interaction between the Coriolis force and the trade winds causes 'Ekman transport', where the surface waters in the tropics move poleward in both hemispheres. Conversely, at latitudes poleward of 25°N and 25°S, the prevailing westerly winds force the surface waters towards the equator, resulting in a convergence of these waters between 15°–30°N and 15°–30°S.[3] This convergence produces a build-up in sea surface height (SSH) around 10°–15°N to 10°–15°S.[4]

The sea-level slope is associated with pressure forces that drive two broad geostrophic westward flows: the North Equatorial Current (NEC) and the South Equatorial Current (SEC) (Figure 2). The NEC and SEC are the equatorial branches of two basin-scale circular circulation patterns: the large subtropical gyres in the northern and southern Pacific Ocean, with their poleward margins outside the domain at higher latitudes.

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[1] Le Borgne, R., Allain, V., Griffiths, S.P., Matear, R.J., McKinnon, A.D., Richardson, A.J. and Young, J.W. 2011. Vulnerability of open ocean food webs in the tropical Pacific to climate change. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 4. pp. 189-250. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[2] Reid, J.L. 1997. On the total geostrophic circulation of the South Pacific Ocean: Flow patterns, tracers and transports. *Progress in Oceanography* 39, 263–352.

[3] Sudre, J. and Morrow, R. 2008. Global surface currents: A high-resolution product for investigating ocean dynamics. *Ocean Dynamics*, doi:10.1007/s10236-008-0134-9

[4] Reid, J.L. 1997. On the total geostrophic circulation of the South Pacific Ocean: Flow patterns, tracers and transports. *Progress in Oceanography* 39, 263–352.

Near the surface (0–80 m), frictional forces from the wind are superimposed on the geostrophic NEC and SEC, adding a component that is perpendicular to the local wind direction due to Ekman transport (see blue streamlines showing Ekman currents in Figure 59). For example, south of the equator the southeasterly trade winds bend the surface streamlines to the southwest.

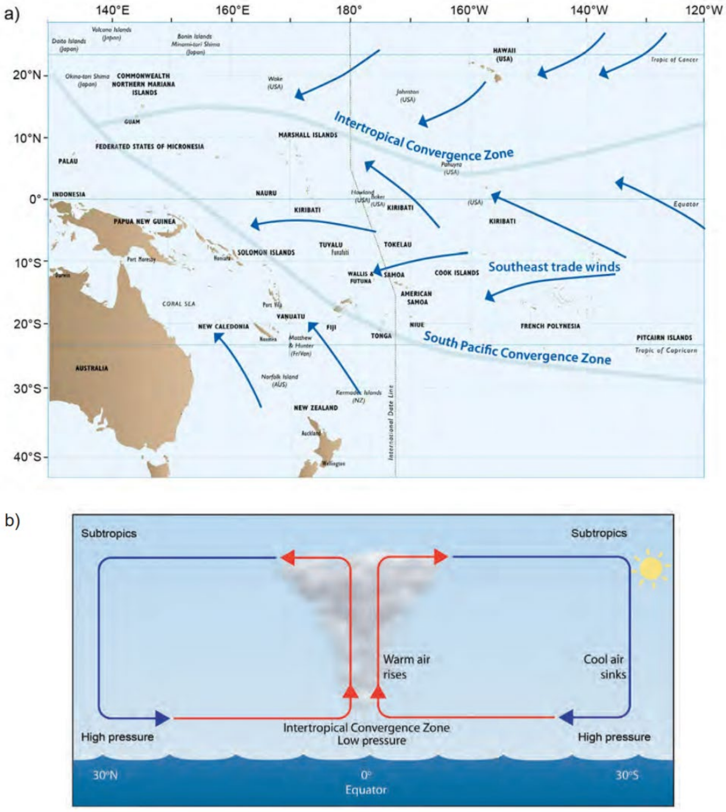

Figure 1. Major atmospheric circulation features in the tropical Pacific with a cross section illustrating the Hadley circulation of the region (b).

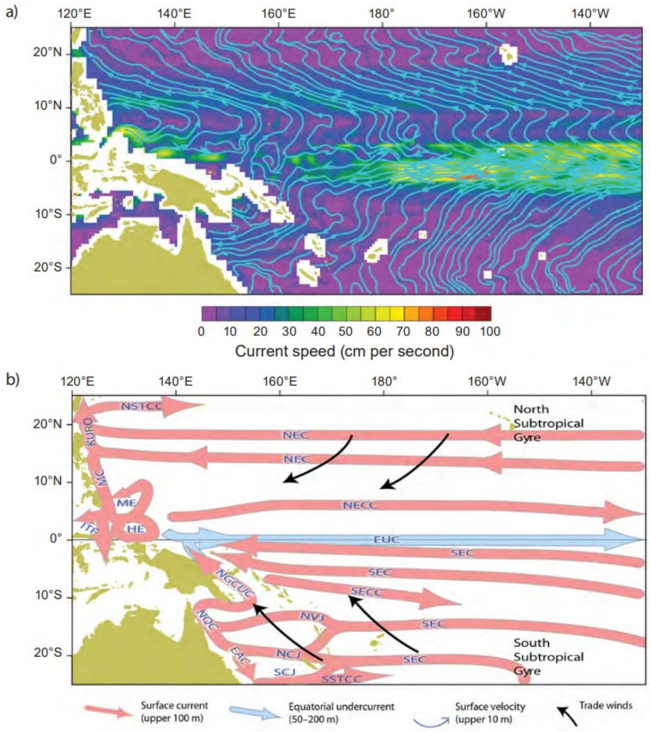

Figure 2. Ocean surface currents in the tropical Western and Central Pacific Ocean, based on satellite data and in situ climatology.[5] (a) As indicated by the streamlines, surface flow is generally directed to the left of the wind in the Southern Hemisphere, and to the right in the Northern Hemisphere, due to Ekman transport. (b) The main ocean currents in the upper 100 to 200 m of the water column. Currents shown are: North Subtropical Counter Current (NSTCC); Kuroshio Current (KURO); Mindanao Current (MC); Mindanao Eddy (ME); Halmahera Eddy (HE); North Equatorial Current (NEC); North Equatorial Counter Current (NECC); Equatorial Undercurrent (EUC); Indonesian Throughflow (ITF); New Guinea Coastal Undercurrent (NGCUC); North Queensland Current (NQC); East Australian Current (EAC); North Vanuatu Jet (NVJ); North Caledonian Jet (NCJ); South Caledonian Jet (SCJ); South Equatorial Counter Current (SECC); South Equatorial Current (SEC) and South Subtropical Counter Current (SSTCC).

Western boundary current system

The westward flowing NEC and SEC both encounter islands in their passage, and eventually the western boundaries of the basin. On average, the NEC occupies a broad region from 8°N to 20°N.[6] As the NEC reaches the western boundary, it divides to feed the Kuroshio Current to the north (called the 'Philippines Current' at this latitude[7]), and the Mindanao Current and Mindanao Eddy to the south (Figure 59). The average latitude where the surface NEC bifurcates is 13.3°N (including Ekman transport) but this varies seasonally from 15°N in July to 17°N in December.[8]

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[5] Sudre, J. and Morrow, R. 2008. Global surface currents: A high-resolution product for investigating ocean dynamics. *Ocean Dynamics*, doi:10.1007/s10236-008-0134-9

[6] Qu, T. and Lukas, R. 2003. The bifurcation of the North Equatorial Current in the Pacific. *Journal of Physical Oceanography* 33, 5–18.

[7] Qu, T. and Lukas, R.. 2003. The bifurcation of the North Equatorial Current in the Pacific. *Journal of Physical Oceanography* 33, 5–18.

[8] Qu, T. and Lukas, R. 2003. The bifurcation of the North Equatorial Current in the Pacific. *Journal of Physical Oceanography* 33, 5–18.

A similar process occurs in the southern Pacific, with the SEC splitting into the North Queensland Current towards the equator and the poleward East Australian Current, with the bifurcation located at 16°S near the surface and 22°S at a depth of 1000 m.[9,10] The location of this bifurcation also varies seasonally by 1.25° latitude, with a displacement toward the equator in November to December.[11]

When these broad currents encounter islands and land masses, the flow is diverted. The pathways of the SEC are complicated by the presence of many islands and ridges in the western Pacific. As a result, the SEC splits into three jets, the South Caledonian Jet, the North Caledonian Jet and the North Vanuatu Jet (Figure 59).[12,13,14]

The northern component of the bifurcation enters the Solomon Sea as the New Guinea Coastal Undercurrent and exits towards the equator through three narrow straits within Solomon Islands.[15] This latter flow, along with the Mindanao Current flow, converges towards the equator, feeding the Warm Pool, the Indonesian Throughflow[16,17] and, at greater depth, the eastward flowing Equatorial Undercurrent (Figure 59).[18]

Eastward flowing counter currents

The westward-flowing NEC and SEC are also altered by the presence of the Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ).[19] These convergence zones alter local wind conditions, generating horizontal shear ('curl') in the wind field. This wind curl produces two eastward flowing, surface-intensified counter currents through Ekman transport and

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[9] Gouriou, Y. and Toole, J. 1993. Mean circulation of the upper layers of the western equatorial Pacific Ocean. Journal of Geophysical Research 98, 22,495–22,520.

[10] Qu, T. and Lindstrom, E. 2002. A climatological interpretation of the circulation in the western South Pacific. *Journal of Physical Oceanography* 32, 2492–2508.

[11] Kessler, W.S. and Gourdeau, L. 2007. The annual cycle of circulation of the south-west subtropical Pacific, analyzed in an ocean GCM. *Journal of Physical Oceanography* 37, 1610–1627.

[12] Webb, D. 2000. Evidence for shallow zonal jets in the South Equatorial Current region of the southwest Pacific. *Journal of Physical Oceanography* 30, 706–720.

[13] Gourdeau, L., Kessler, W.S., Davis, R.E., Sherman, J. and others. 2008. Zonal jets entering the Coral Sea. *Journal of Physical Oceanography* 38, 715–725.

[14] Ganachaud, A., Gourdeau, L. and Kessler, W. 2008. Bifurcation of the subtropical south equatorial current against New Caledonia in December 2004 from a hydrographic inverse box model. *Journal of Physical Oceanography* 38, 2072–2084.

[15] Lindstrom, E., Lukas, R., Rine, R., Firing, E. and others. 1987. The western equatorial Pacific Ocean circulation study. *Nature* 330, 533–537

[16] Gordon, A. and Fine, R. 1996. Pathways of water between the Pacific and Indian oceans in the Indonesian seas. *Nature* 379, 146–149.

[17] Wijffels, S., Meyers, G. and Godfrey, S. 2008. A 20-year average of the Indonesian throughflow: Regional currents and the inter-basin exchange. *Journal of Physical Oceanography* 38, 1965–1978.

[18] Fine, R., Lukas, R., Bingham, F.M., Warner, M.J. and Gammon, R.H. 1994. The western equatorial Pacific is a water mass crossroads. *Journal of Geophysical Research* 99, 25,063–25,080.

[19] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

geostrophic processes, known as the North Equatorial Counter Current (NECC) under the ITCZ[20], and the South Equatorial Counter Current (SECC) under the SPCZ (Figure 3). [21,22]

Another modification to the flow comes from local changes in the trade winds as they interact with island topography. This interaction modifies the flow of water downwind of islands. In the southern Pacific, similar counter currents occur in the Coral Sea west of Vanuatu[23] and near the Marquesas Islands[24,25] causing the upwelling of nutrients. Near Hawaii, an eastward flowing current is created that flows against the NEC, named the Hawaii Lee Counter Current.[26] Eddies are formed at the boundaries where the westward flowing NEC and the eastward flowing NECC meet. These eddies also have the potential to bring nutrients to the surface waters.[27]

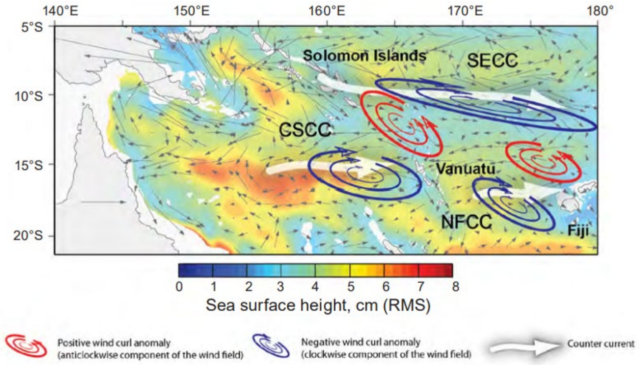

Figure 3. Formation of counter currents in the tropical Pacific Ocean.[28] Weakening trade winds towards the north of the South Pacific Convergence Zone area create a clockwise forcing on the ocean by the wind field (negative wind curl). This generates pressure forces in the ocean that set-in motion a counter current, the South Equatorial Counter Current (SECC) that flows against the broad westward South Equatorial Current. West of Vanuatu and Fiji, similar disruptions to the trade winds create wind stress curl dipoles that generate the Coral Sea Counter Current (CSCC) and the North Fiji Counter Current (NFCC). The presence of counter currents is also revealed by variability in sea surface height, as measured by satellite data (colour) over short time scales (180 days or less). RMS = root mean square of sea surface height, in cm.

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[20] Johnson, G., Sloyan, B., Kessler, W. and McTaggart, K. 2002. Direct measurements of upper ocean currents and water properties across the tropical Pacific during the 1990s. *Progress in Oceanography* 52, 31–61.

[21] Gouriou, Y. and Toole, J. 1993. Mean circulation of the upper layers of the western equatorial Pacific Ocean. *Journal of Geophysical Research* 98, 22,495–22,520.

[22] Reid, J.L. 1959. Evidence of a South Equatorial Counter-current in the Pacific Ocean. *Nature* 184, 209–210.

[23] Qiu, B., Chen, S. and Kessler, W. 2009. Source of the 70-day mesoscale eddy variability in the Coral Sea and the North Fiji Basin. *Journal of Physical Oceanography* 39, 404–420.

[24] Martinez, E. and Maamaatuaiahutapu, K. 2004. Island mass effect in the Marquesas Islands: Time variation. *Geophysical Research Letters* 31, doi:10.1029/2004GL020682

[25] Martinez, E., Ganachaud, A., Lefevre, J. and Maamaatuaiahutapu, K. 2009. Central South Pacific thermocline water circulation from a high-resolution ocean model validated against satellite data: Seasonal variability and El Niño 1997–1998 influence. *Journal of Geophysical Research* 114, C05012, doi:10.1029/2008JC004824

[26] Qiu, B., Koh, D., Lumpkin, C. and Flament, P. 1997. Existence and formation mechanism of the North Hawaiian Ridge Current. *Journal of Physical Oceanography* 27, 431–444.

[27] Calil, P.H.R., Richards, K.J., Yanli J. and Bidigare, R.R. 2008. Eddy activity in the lee of the Hawaiian Islands. *Deep-Sea Research II* 55, 1179–1194.

[28] Qiu, B., Chen, S. and Kessler, W. 2009. Source of the 70-day mesoscale eddy variability in the Coral Sea and the North Fiji Basin. *Journal of Physical Oceanography* 39, 404–420.

Temperature

Water temperature is a key feature of the ocean that affects the abundance and distribution of marine life supporting fisheries in the oceans. Increases in temperature above normal maxima are expected to have negative effects on the overall viability of some populations of fish and invertebrates.[29] The effects of water temperature on the distribution and abundance of skipjack tuna are of special interest to Pacific Island countries considering the strong influence that ENSO events have on the distribution of this species.[30]

The SST of the tropical Pacific Ocean varies spatially and temporally. More solar heat is absorbed by the ocean near the equator than at higher latitudes. This rise in temperature results in a pole-to-equator SST gradient.[31] However, there are regional deviations from this large-scale pattern. For example, the southeast trade winds push the warmest waters to the western side of the Pacific basin, forming the Warm Pool – a large heat reservoir intimately associated with ENSO.

Along the equator and the coastline of South America, the prevailing winds cause equatorial and coastal upwelling, respectively, bringing cool, deep waters rich in nutrients to the surface. This process results in a relatively cool tongue of water extending from South America along the equator to the central Pacific, where it meets the eastern edge of the Warm Pool near the dateline.

Sea surface temperature also varies seasonally. Away from the equator, SST varies by up to 7°C throughout the year; whereas seasonal changes in SST near the equator are weak and the largest variations (2 to 3°C) occur from one year to another. This interannual variability is mainly associated with ENSO.[32] Thus, SST at any given location and time depends on a broad range of processes, including diurnal and seasonal solar heating modulated by cloudiness, air-sea heat exchanges, heat transport by oceanic circulation, eddies and other local ocean processes, such as vertical mixing and upwelling (Figure 4).

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[29] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[30] Lehodey, P., Hampton, J., Brill, R.W., Nicol, S., Inna Senina, I., Calmettes, B., Pörtner, H.O., Bopp, L., Ilyina, T., Bell, J.D. and Sibert, J. 2011. Vulnerability of oceanic fisheries in the tropical Pacific to climate change. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 8. pp 433-491. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[31] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages. Already listed

[32] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

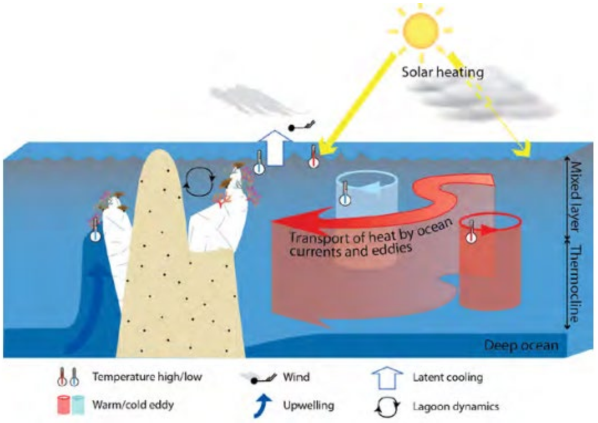

Figure 4. Factors affecting the temperature of the upper layer of the ocean. Surface waters are heated by incoming solar radiation, modulated by cloud cover. Winds create increased evaporation, which extracts latent heat from the ocean causing it to cool. Contact between the atmosphere and ocean also drives direct heat fluxes. The wind generates mixing with deeper and colder waters, resulting in cooling at the surface and more homogenised temperatures within the 'mixed layer'. Currents, eddies and upwellings also transport warmer or colder water into or away from the upper layer of the ocean.[33]

Vertical temperature structure

The temperature of the tropical Pacific Ocean declines as depth increases (Figure 5). This variation in temperature occurs because most of the sun's heat is absorbed near the surface, with little energy reaching below the first 100 m. The warmer surface water has lower density than the deeper cooler waters below and where these two layers meet at the 'thermocline', the water temperature changes rapidly. In the tropical Pacific Ocean, the thermocline usually lies within the upper 500 m of the water column and the temperature drops by about 20°C. Below this layer, the ocean remains cold to abyssal depths.

The stratification across the tropical Pacific is generally strong but decreases with increasing latitude (Figure 6). The strongest stratification occurs below the main atmospheric convergence zones (ITCZ and SPCZ).[34]

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[33] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[34] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

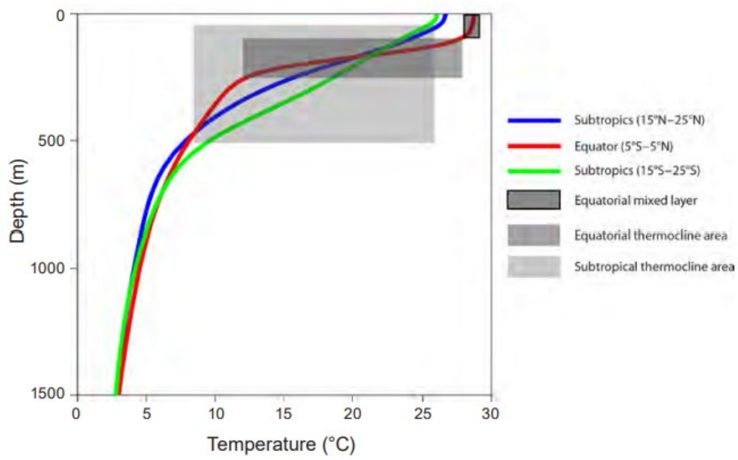

Figure 5. Average water temperature from the surface to a depth of 1500 m for three typical locations in the tropical Pacific : equator (red line); north subtropics blue line); and south subtropics (green line). The equatorial thermocline area is indicated by the darker grey bar; the lighter grey bar indicates the area of the thermocline in the subtropics. All profiles are averaged between 160°E and 160°W.[35] Because water density decreases with increasing temperature and increases with increasing salinity (but to a lesser extent), 'stratification' occurs in the water column. That is, lighter surface water remains separated from the denser deeper layer, like a layer of oil sitting on water. Increased stratification makes the water column more stable because more energy, in the form of wind or buoyancy-driven convection, is needed to mix water between the two layers. This requirement has important implications for fisheries because when the water column is stable the transfer of nutrients to the sunlit (photic) zone where primary production occurs is inhibited.[36]

Nevertheless, several processes occur in the surface layer of the ocean which can act against stratification. In addition to generating surface currents, wind causes strong mixing in the upper layer. Evaporation, and loss of heat from surface waters at night or during winter at higher latitudes, also causes strong convective mixing in the upper layer and weakens the stratification.

These forms of turbulence and mixing homogenise temperature, salinity and other properties of the ocean in the first few tens of metres (Figure 7), resulting in the surface 'mixed layer'. This homogenisation of properties results in a very sharp gradient at the base of the mixed layer. Below this level, the properties of the water column are affected by deeper oceanic processes. The depth of the mixed layer varies substantially on daily to interannual time scales and ranges from about 0 to 200 m at low latitudes. In nutrient-poor (oligotrophic) waters, the mixed layer depth is vitally important because it determines the depth from which nutrients are brought up to support biological activity at the surface. [37,38]

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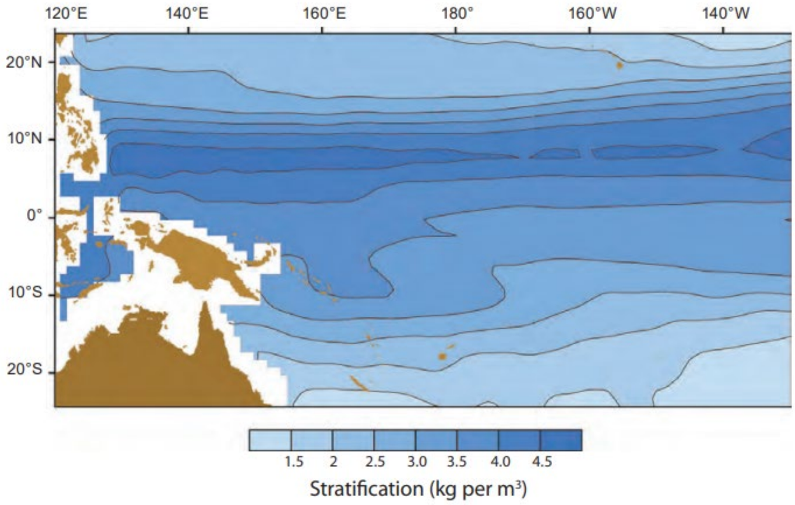
[35] Ridgway, K.R. and Dunn, J.R. 2003. Mesoscale structure of the East Australian Current System and its relationship with topography. *Progress in Oceanography* 56, 189–222.

[36] Ganachaud, A.S., Sen Gupta, A., Orr, J.C., Wijffels, S.E., Ridgway, K.R., Hemer, M.A., Maes, C., Steinberg, C.R., Tribollet, A.D., Qiu, B. and Kruger, J.C. 2011. Observed and expected changes to the tropical Pacific Ocean. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 3. pp. 101-188. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[37] Holbrook, N.J. and Bindoff, N.L. 1999. Seasonal temperature variability in the upper southwest Pacific Ocean. *Journal of Physical Oceanography* 29, 366–381.

[38] De Boyer Montégut, C., Madec, G., Fischer, A., Lazar, A. and Iudicone, D. 2004. An examination of profile data and a profile-based climatology. *Journal of Geophysical Research* 109, C12003, doi:10.1029/2004JC002378.



[39] Stratification is defined here as the difference in water density (kg per m^3) between depths of 13 m and 200 m (13 m was chosen for compatibility with the common upper level of IPCC-AR4 model analyses). The highest vertical stratification is found under the Intertropical Convergence Zone and the South Pacific Convergence Zone.

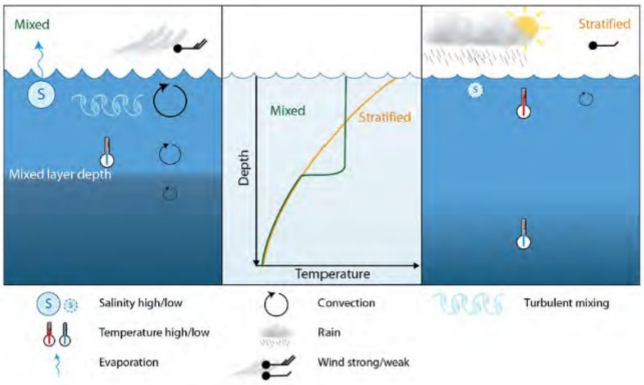


Figure 7. Factors affecting the mixed layer depth. The left panel shows a well-formed mixed layer with homogenised properties in the upper part of the water column; the right panel shows a smoothly stratified field. The corresponding temperature profiles are given in the middle panel. The formation of the mixed layer is enhanced by cooling, evaporation and strong winds, and weakened by heating, precipitation and slight winds.[40]

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[39] Carton, J.A., Chepurin, G.A., Cao, X. and Giese, B. 2000. A simple ocean data assimilation retrospective analysis of the global ocean 1950–1995. Part I: Methodology. *Journal of Physical Oceanography* 30, 294–309.

[40] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

In the upper ocean, two warm water 'bowls' appear between 10°–20°N and 10°–20°S and extend to depths of 500 m (Figure 8). These bowls are formed by the convergence of surface waters under the influence of Ekman transport; the easterly trade winds drive the flow of warm surface water in the tropics poleward, whereas at 40°N–40°S westerly winds drive surface water towards the equator. This leads to a convergence that pushes the surface water downwards and creates the bowls of warmer water.

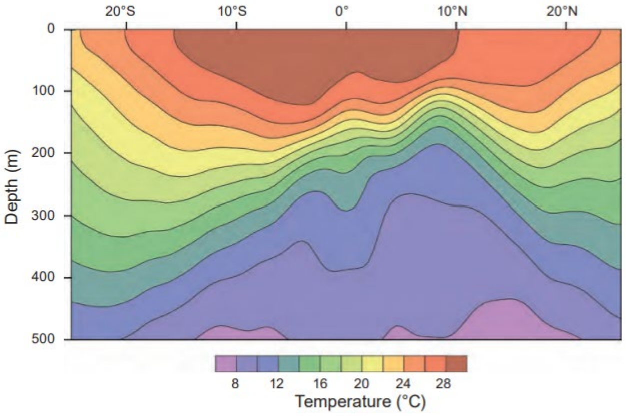

Figure 8. Vertical temperature structure (°C) of the water column , to a depth of 500 m, from 25°S to 25°N (averaged across the area from 160°E to 160°W). Note that warmer surface water penetrates to greater depths between 15°–20°N and 15°–20°S.[41]

In many parts of the tropical Pacific, the most important factor affecting water density, and therefore stratification, is temperature. However, in some regions, such as the Warm Pool, salinity becomes equally important. The effects of temperature and salinity changes on density are significant because the motion of the ocean is controlled largely by spatial differences in water density. For example, the downward tilt (towards the north) in the temperature structure below 100 m between 10° and 15°N (Figure 8) causes a westward flowing geostrophic current – the NEC. Similarly, the upward tilt (towards the north) in the deeper isotherms of the Southern Hemisphere is associated with the westward flowing, geostrophic SEC.

The Warm Pool

The Western Pacific Warm Pool Province[42], is of major significance to tuna fisheries in the tropical Pacific.[43] The combined Warm Pool waters of the equatorial western Pacific Ocean and equatorial eastern Indian Ocean cover a vast area characterised by the world's warmest oceanic temperatures, the deepest atmospheric convection and the heaviest precipitation.[44]

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[41] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[42] Longhurst, A.R. 2006. *Ecological Geography of the Sea*. Academic Press, New York, United States of America.

[43] Le Borgne, R., Allain, V., Griffiths, S.P., Matear, R.J., McKinnon, A.D., Richardson, A.J. and Young, J.W. 2011. Vulnerability of open ocean food webs in the tropical Pacific to climate change. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 4. pp. 189-250. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[44] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

This Indo-Pacific Warm Pool is often defined by SSTs greater than a threshold value – usually 28°C.[45] By this definition, it extends over ~ 12 million km^2 of the tropical Pacific Ocean and ~ 6 million km^2 of the Indian Ocean. The interactions between the ocean and the atmosphere in this part of the world occur through radiative and latent heat transfers that drive vigorous convective clouds, winds and precipitation.[46] Small changes at the ocean-atmosphere boundary are enhanced through convection and can lead to self-amplification and eventually large changes in atmospheric circulation. The high precipitation levels above the Warm Pool strongly affect salinity, resulting in a relatively warm and fresh (less dense) pool of well-mixed water overlaying a cold and relatively saline (denser) ocean interior (Figure 9).

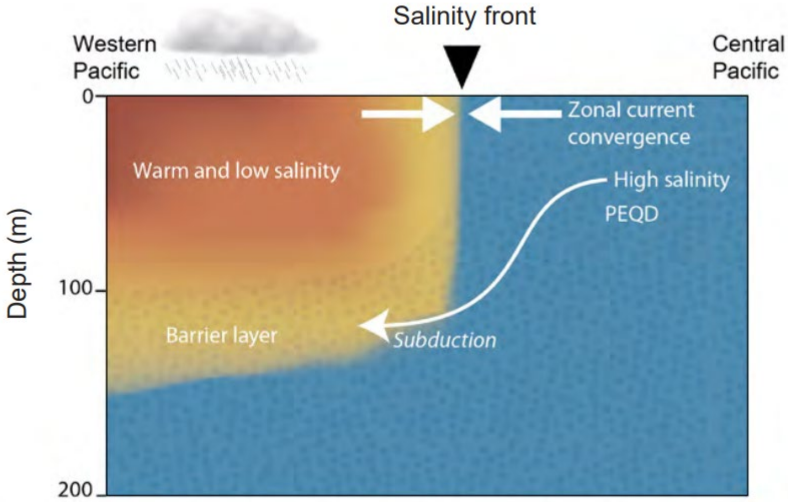

Figure 9. Temperature structure of the equatorial Pacific showing the Warm Pool. The thermally homogeneous waters of the Warm Pool, freshened by heavy rainfall in the western Pacific, converge with the colder, saltier waters originating in the Pacific Equatorial Divergence Province (PEQD) to the east. This convergence is one of the mechanisms that leads to the formation of a salinity-stratified layer in the lower part of the Warm Pool known as the 'barrier layer'. Upwelling occurs to the east of the salinity front, bringing nutrient-rich waters to the surface. In contrast, the Warm Pool has low nutrient levels.[47]

The Warm Pool is at the heart of the ENSO mechanism.[48] Under 'normal' conditions, the trade winds generate surface currents that push and accumulate warm water to the west[49], creating a large heat reservoir that maintains the easterly trade winds via the Walker circulation.[50]

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[45] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[46] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[47] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[48] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 2. pp. 49-100. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[49] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

[50] Lough, J.M., Meehl, G.A. and Salinger, M.J. 2011. Observed and projected changes in surface climate of the tropical Pacific. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and*

When an El Niño event occurs, the trade winds weaken, allowing the Warm Pool to spread eastwards across the equatorial Pacific over a period of ~ 2 months, along with its wind and precipitation systems. The displacement of the Warm Pool occurs through wave dynamics that set the speed and structure of the Warm Pool's motion to the east. After an El Niño event, westward propagating waves reset the system towards a 'normal' situation, replenishing and relocating the Warm Pool. Sometimes an 'overshoot' gives rise to a period of more extreme trade winds and a westward contraction of the Warm Pool – the Pacific shifts to a La Niña situation. El Niño events are associated with a 'shoaling' of the Warm Pool thermocline, which enhances primary productivity.[51] Conversely, the thermocline deepens during La Niña episodes, leading to lower productivity. At the eastern edge of the Warm Pool, there is a convergence zone where the incoming westward SEC weakens and there is a small eastward flow within the Warm Pool (Figure 10).

The interannual movements of the Warm Pool (and its eastern edge) are controlled by the relative strength of these zonal currents and are subject to variations that are in phase with ENSO.[52] Because of the contrasting water properties to the east and west of this boundary, the convergence zone is characterised by a sharp change in salinity (Figure 10).

Displacements of this convergence zone are an intrinsic part of the ENSO system[53] and can reach several thousands of kilometres, between 140°W during El Niño conditions and as far west as 140°E during La Niña episodes.[54] These displacements also determine the boundary between the Pacific Equatorial Divergence Province (PEQD) and the Warm Pool.[55]

The relative contributions of temperature and salinity to the density of sea water are similar in the Warm Pool – a situation that is uncommon in most other parts of the tropical Pacific Ocean, where temperature normally dominates. On average, temperature is homogeneous in the upper 60 m of the water column in the Warm Pool with a sharp thermocline below this surface layer. In contrast, the salinity 'halocline' in the Warm Pool is much shallower, resulting in a mixed layer (where both temperature and salinity are well-mixed), which is only about 30 m deep.[56] The stratified layer between the thermocline and the base of the mixed layer is called the 'barrier layer' (Figure 10). Within the barrier layer regions, any mixing driven by the wind is restricted so that the waters below the barrier layer are insulated from the influence of the atmosphere, and vice versa. This stratification acts as a barrier to the vertical exchanges of heat, fresh water and nutrients, and

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[51] Mackay, D.J., O'Sullivan, J.E. and Watson, R.J. 2002. Iron in the western Pacific: A riverine or hydrothermal source for iron in the Equatorial Undercurrent? *Deep-Sea Research* I 49, 877–893.

[52] Picaut, J., Ioualalen, M., Menkes, C., Delcroix, T. and McPhaden, M.J. 1996. Mechanism of the zonal displacements of the Pacific warm pool: Implications for ENSO. *Science* 274, 1486–1489.

[53] Picaut, J., Masia, F. and du Penhoat, Y. 1997. An advective-reflective conceptual model for the oscillatory nature of the ENSO. *Science* 277, 663–666.

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suppresses mixing.[57,58] The location and strength of the barrier layer are important to both ENSO dynamics[59,60] and biological production.[61]

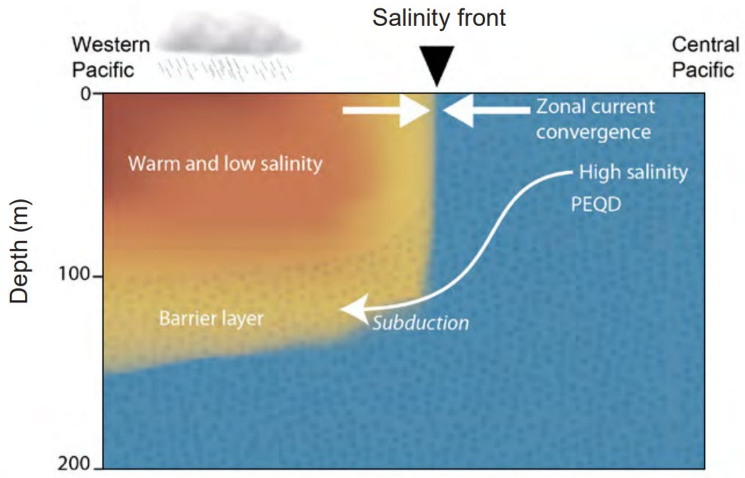

Figure 10. Temperature structure of the equatorial Pacific showing the Warm Pool. The thermally homogeneous waters of the Warm Pool, freshened by heavy rainfall in the western Pacific, converge with the colder, saltier waters originating in the Pacific Equatorial Divergence Province (PEQD) to the east. This convergence is one of the mechanisms that leads to the formation of a salinity-stratified layer in the lower part of the Warm Pool known as the 'barrier layer'. Upwelling occurs to the east of the salinity front, bringing nutrient-rich waters to the surface. In contrast, the Warm Pool has low nutrient levels.[62]

Sea level

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[57] Eldin, G., Delcroix, T. and Rodier, M. 2004. The frontal area at the eastern edge of the western equatorial Pacific warm pool in April 2001. *Journal of Geophysical Research* 109, doi:10.1029/2003JC002088.

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[60] Maes, C., Picaut, J. and Belamari, S. 2002. Salinity barrier layer and onset of El Niño in a Pacific coupled model. *Geophysical Research Letters* 29, 2206, doi:10.1029/2002GL016029.

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[62] Sen Gupta A., Santoso, A., Taschetto, A., Ummenhofer, C.C. and others. 2009. Projected changes to the Southern Hemisphere Ocean and sea ice in the IPCC-AR4 climate models. *Journal of Climate* 22, 3047–3078.

Sea level components and variations

Sea level at a given location is determined by a number of factors (Figure 11), and the interaction of these factors causes substantial variation on a broad range of time scales (Figure 11). Tides affect sea level on a predictable periodic basis; storms and eddies are episodic with effects lasting from hours to days; and circulation changes, like those associated with ENSO, can cause large year-to-year variation.

The steady driver of global sea-level rise, however, is related to the long-term warming of the ocean and atmosphere which causes the melting of land-based ice and thermal expansion of sea water. Both processes increase the volume of the ocean. During the peak of the last ice age, when ice sheets covered large parts of the Northern Hemisphere, the additional storage of water on the land led to a drop in sea level of ~ 120 m. Conversely, during the Pliocene ~ 3 million years ago, when global average temperatures were thought to be 2 to 3°C above today's temperatures, sea level was at least 15 to 20 m above present-day values.[63]

Today, the main ice sheets are in Greenland and Antarctica, supplemented by ice caps (small ice sheets) such as the one in Iceland, and glaciers. If the major ice sheets were to melt completely, sea level would increase by about 70 m.[64] In contrast, complete melting of the ice caps and glaciers would increase sea level by only about 70 cm. The melting of floating sea ice does not affect sea level. Regional changes in sea level due to ENSO events can be as great as 20 to 30 cm (Figure 11).[65,66]

The high degree of natural variability makes it difficult to extract robust long-term trends from the relatively short-term records for the region. Combining the information from individual tide gauges to produce estimates for the entire tropical Pacific requires advanced statistical methods[67], particularly because of differences in the way ENSO affects sea level in different locations.

In general, an El Niño episode will tend to lower the sea level in the west, and raise it in the east, along the equator. In contrast, a La Niña event can significantly raise the sea level to the west. Over long time periods, sea-level rise due to El Niño events balances the fall due to La Niña episodes. Based on the long-term data series from tide gauges in the tropical Pacific, sea level is rising by between 0 and 3 cm per decade, depending on location (Figure 12).

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[65] Church, J., White, N. and Hunter, J. 2006. Sea-level rise at tropical Pacific and Indian Ocean islands. *Global and Planetary Change* 53, 155–168.

[66] And see: https://nsidc.org/arcticseaicenews/#:~:text=On%20September%2010%2C%202023%2C%20sea,in%201979%20(Figure%201).

[67] Church, J.A., White, N.J., Coleman, R., Lambeck, K. and Mitrovica, J.X. 2004. Estimates of the regional distribution of sea-level rise over the 1950 to 2000 period. *Journal of Climate* 17, 2609–2625.

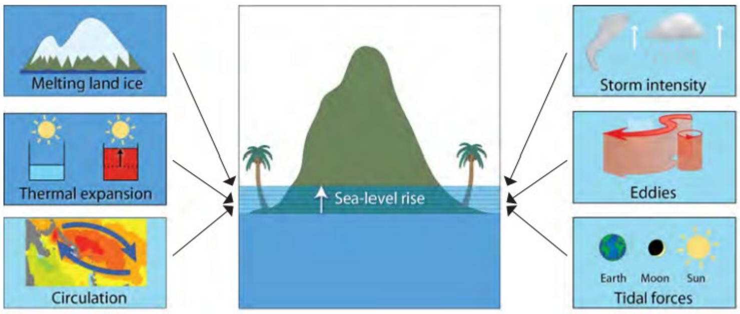

Figure 11. Factors determining sea-level rise at any given time and location: melting of land ice, thermal expansion of the ocean, large-scale ocean circulation, atmospheric pressure and storm surges, transient eddies (or ENSO-type waves at the equator) and tides. Melting of land ice and thermal expansion of the ocean are the most important of these factors.[68]

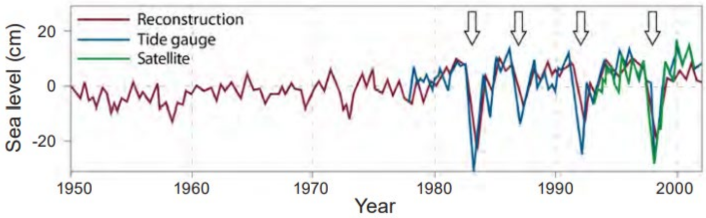

Figure 12. Time series (1950–2004) from Funafuti, Tuvalu, showing variation in sea level. Data are from tide gauge (blue), satellite (green) and reconstructions based on the empirical relationship between spatial patterns of variations and tide gauge data (red). Major drops in sea level occur during El Niño events (see arrows). A gradual long-term upward trend can be seen which is particularly pronounced in the second half of the record.[69]

Changes in large-scale currents can also alter sea level. An acceleration of the subtropical gyre in the Southern Hemisphere due to long-term changes in the wind field has led to modified sea levels, with local increases or decreases of ~ 10 cm over several years.[70] Oceanic eddies can alter sea level by ~ 10 cm for several days.[71] Decreases in atmospheric pressure raise sea level by 1 cm for each millibar decrease (Figure 13). This effect is most extreme during cyclones, when the low atmospheric pressure at their centre can draw the water column up by 30 to 50 cm. The amplitudes of each process affecting sea level are usually smaller than those of tides, but they can act in

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[68] Ganachaud, A.S., Sen Gupta, A., Orr, J.C., Wijffels, S.E., Ridgway, K.R., Hemer, M.A., Maes, C., Steinberg, C.R., Tribollet, A.D., Qiu, B. and Kruger, J.C. 2011. Observed and expected changes to the tropical Pacific Ocean. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 3. pp. 101-188. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

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concert. For example, storm surges associated with higher sea levels during cyclones allow destructive waves to penetrate further into coastal habitats.[72]

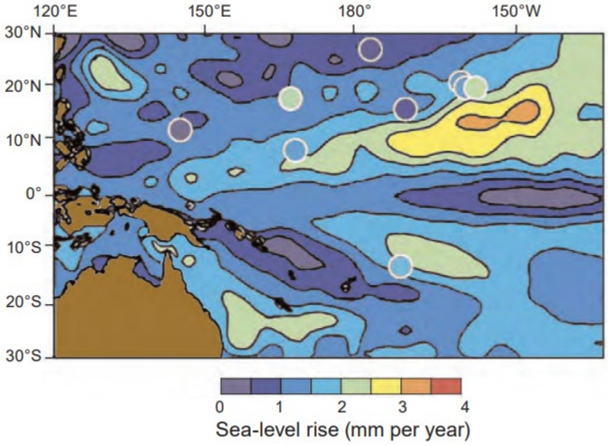

Figure 13. Sea level trends from 1950–2001, reconstructed from tide gauge data at several locations across the tropical Pacific Ocean (coloured dots). Recent satellite data have been used to determine the relationship between localised tide gauge data and the rest of the ocean. The longer time-series of tide gauge data are then projected over the region.[73]

Nutrient supply

The availability of nutrients in the ocean is of great significance to fisheries. Together with sunlight, nutrients underpin the primary productivity associated with phytoplankton at the base of the food webs that support the stocks of tuna and associated oceanic pelagic fish harvested throughout the tropical Pacific Ocean. The main nutrients associated with biological productivity are nitrates, phosphates and silicates. The maintenance of this productivity can be fragile, however, because nutrients are not distributed evenly – they are depleted near the surface, where they are needed, but abundant in the deeper ocean. This variation occurs because the phytoplankton use up the available nutrients in the photic zone, where there is sufficient light for photosynthesis and, although a small part of the nutrients pass down the food web, most of them eventually sink as organic matter into the deep ocean. There, bacteria remineralise the organic matter, releasing nutrients. As a result, concentrations of nutrients are much greater below the pycnocline than they are at the surface (Figure 14).

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[72] Ganachaud, A.S., Sen Gupta, A., Orr, J.C., Wijffels, S.E., Ridgway, K.R., Hemer, M.A., Maes, C., Steinberg, C.R., Tribollet, A.D., Qiu, B. and Kruger, J.C. 2011. Observed and expected changes to the tropical Pacific Ocean. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 3. pp. 101-188. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

[73] Church, J., White, N. and Hunter, J. 2006. Sea-level rise at tropical Pacific and Indian Ocean islands. *Global and Planetary Change* 53, 155–168.

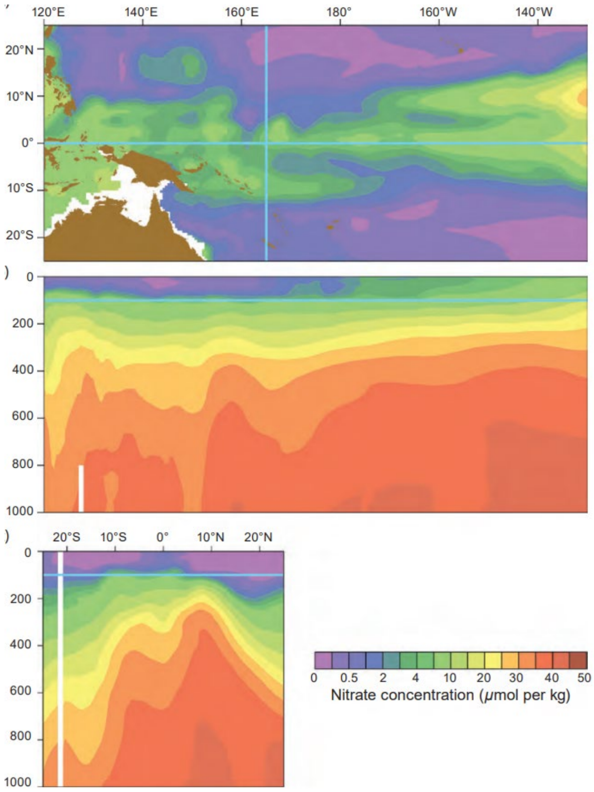

Figure 14. (a) Average dissolved nitrate concentration (μmol per kg) in the tropical Pacific Ocean at a depth of 100 m; and average dissolved nitrate concentrations for vertical sections of the water column to depths of 1000 m at (b) the equator, and (c) 165°E. The positions of the vertical sections are indicated as blue lines on (a); horizontal blue lines in (b) and (c) represent values at a depth of 100 m as shown in (a).[74] White areas in (b) and (c) correspond to topography.

The sinking of organic material is also one of the main ways that CO_2 is transferred into the deep ocean from the atmosphere and is known as the 'biological pump'.[75] Ocean circulation, or deep mixing of the water column, is needed to transfer the nutrients back to the surface layers. The strong density stratification (pycnocline) usually associated with the thermocline effectively inhibits the vertical exchange of water (and therefore nutrients) between the deep and shallow layers of the ocean. Molecular mixing is too weak to transport significant amounts of nutrients towards the surface. The main processes that can overcome the stratification barrier and deliver nutrient-rich water to the upper layers are turbulence in the mixed layer, wind-driven upwelling and eddies.

- ⇨ Turbulence in the mixed layer: The strong mixing in the upper ocean entrains nutrient-rich deeper waters and circulates them within the photic zone. The extent of nutrient input is determined by the depth of mixing in relation to the depth of the nutricline. A strong wind, for example, may deepen the mixed layer to 100 m in several hours and entrain the nutrient-rich deep waters from greater depths. On seasonal time scales, and at subtropical latitudes, the mixed layer deepens during winter because of stronger winds and surface cooling and becomes shallower during summer as surface warming acts to stratify the water column (Figure 15).

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- ⇨ Upwelling: Upwelling of deep, nutrient-rich waters is created by the divergence of surface water, when wind-driven surface currents move in opposite directions (e.g. at the equator), or when surface water is pushed away from the coast (e.g. near islands, or along the Peru-Chile coast). The surface divergence 'draws' deeper waters towards the surface, creating an 'upwelling'. If these waters originate from a sufficient depth, they deliver nutrients into the mixed layer. This occurs in PEQD, where strong upwelling occurs within 4 to 5° of the equator driven by the easterly trade winds.[76] In contrast, the subtropics (NPTG and SPSG) are dominated by convergence of surface waters and downwelling, making them areas of low productivity. Weak upwelling is found, however, near the poleward edge of the SPCZ and ITCZ, and near islands in NPTG and SPSG, where sharp changes in the direction of ocean currents or the wind can lead to local upwelling events that enrich surface waters. The local influence of coasts and islands is more pronounced in ARCH, where boundary currents, jets, wind-driven upwelling, enhanced internal waves or tidal mixing activity more commonly bring nutrients into the photic zone.[77]

- ⇨ Eddies: Surface divergence associated with oceanic eddies and Rossby waves, both within the eddies or waves (mesoscale) and at their edges (sub mesoscale), can raise the thermocline temporarily and bring nutrient-rich waters into the photic zone to increase primary productivity.[78,79,80,81] Because oceanic eddies are ubiquitous in the subtropics (Figure 17), they are believed to be the main nutrient supply system along with frontal processes in NPTG and SPSG.[82,83, 84,85,86,87]

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[76] Le Borgne, R., Allain, V., Griffiths, S.P., Matear, R.J., McKinnon, A.D., Richardson, A.J. and Young, J.W. 2011. Vulnerability of open ocean food webs in the tropical Pacific to climate change. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 4. pp. 189-250. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

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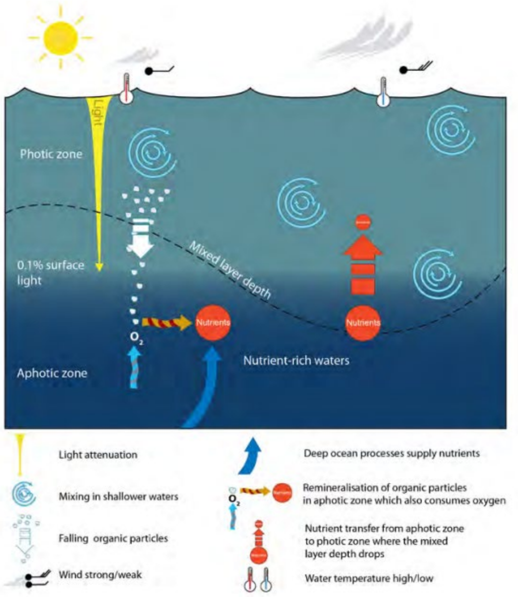

Figure 15. Factors contributing to nutrient concentration in the surface waters of the tropical Pacific Ocean. A deeper mixed layer, caused by stronger winds and lower surface temperatures during winter, helps to transfer nutrients to the photic zone, where they are available to contribute to biological production. Advection by ocean currents and decomposition of organic matter by bacteria (remineralisation) maintains a reservoir of nutrients below the mixed layer.[88]

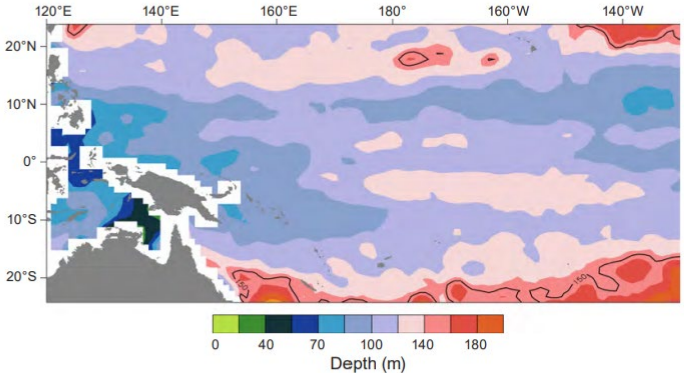

Figure 16. Maximum mixed layer depths across the tropical Pacific Ocean for the period 1980–1999, based on the Simple Ocean Data Assimilation (SODA).[89] The mixed layer depth (MLD) is defined as the depth at which the density of the water increases by 0.1 kg per m3 with respect to the density at the surface.[90] The 150 m MLD contour is shown in black.

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[88] Ganachaud, A.S., Sen Gupta, A., Orr, J.C., Wijffels, S.E., Ridgway, K.R., Hemer, M.A., Maes, C., Steinberg, C.R., Tribollet, A.D., Qiu, B. and Kruger, J.C. 2011. Observed and expected changes to the tropical Pacific Ocean. In: Bell J.D., Johnson, J.E. and Hobday, A.J. 2011. *Vulnerability of Tropical Pacific Fisheries and Aquaculture to Climate Change*. Chapter 3. pp. 101-188. Secretariat of the Pacific Community, Noumea, New Caledonia. 941 pages.

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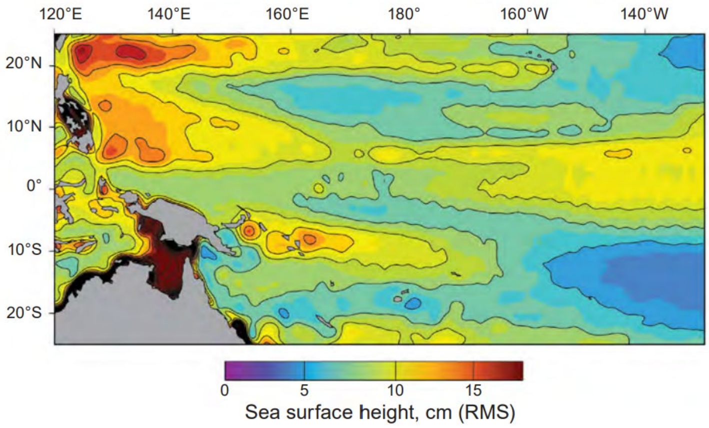

Figure 17. Eddies in the tropical Pacific Ocean, shown as variability (standard deviation) in sea surface height (SSH) anomaly from a merged satellite altimetry dataset (including Topex/Poseidon, Jason, ERS and Envisat from 1992–2007). Contour interval is 2.5 cm.[91] Note that passing eddies cause a temporary change in sea level.

Overall, the supply of nutrients to the photic zone in the tropical Pacific Ocean is the result of the interplay between mixed layer depth, upwelling and eddies against the background of the stratification and nutrient structure of the particular ocean province. Upwelling, for example, driven by the easterly trade winds, plays an important part in supplying nutrients in PEQD, where the thermocline is shallow.

In addition, ENSO-related equatorial waves can lead to displacements in the thermocline, which affect the supply of nutrients. Following a westerly wind burst in the western Pacific (an important precursor to El Niño events), a ‘downwelling’ equatorial wave moves across the basin to the east deepening the thermocline on its way and suppressing upwelling. The strength and depth of the EUC is another important factor controlling productivity in the tropical eastern Pacific as it provides a source of iron, which is the element that limits primary production in that region.[92,93,94] The sporadic nature of phytoplankton blooms in the tropical Pacific Ocean suggests that occurrence and coincidence of the various processes affecting biological production along the equator are highly

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irregular.[95] A single biological bloom associated with a La Niña episode can dominate the average conditions over a 5-year period.[96]

Dissolved oxygen

Adequate levels of dissolved oxygen (O_2) throughout the water column are essential for the growth and survival of the zooplankton and micronekton in the food webs for tuna and other oceanic pelagic fish, and for tuna and associated pelagics themselves.[97,98] Where the concentrations of dissolved oxygen are too low, the distributions of these animals are restricted because they require O_2 for energy, and for oxidising organic substrates, such as carbohydrates.

The levels of dissolved oxygen in surface waters are determined by the rate at which oxygen is transferred from the atmosphere (which is highly dependent on SST and surface mixing), the rate it is produced from photosynthesis by phytoplankton, and the rate at which the oxygen-rich surface waters are submerged via ocean currents and mixing. At high latitudes, some cold surface waters rich in O_2 are pushed dynamically to lower latitudes, below lighter subtropical waters – a process called 'subduction'. These waters gradually lose O_2 as it is used up in the decomposition of organic matter by bacteria, a process termed 'remineralisation'. Therefore, dissolved oxygen concentration at any point in the water column is a balance between the original O_2 content, the effect of remineralisation of organic matter, and the rate at which water is replaced through ocean circulation. As a result, vertical O_2 concentration is usually inversely related to nutrient levels in the upper 500 m.[99]

In regions of high remineralisation, consumption of O_2 can exceed replenishment from ocean circulation, causing part of the water column to become depleted in oxygen. This depletion results in anoxic conditions (see for example the areas around 10°N and 5°S (Figure 18). Unlike CO_2, changes in the dissolved oxygen content in surface waters are insensitive to changes in atmospheric O_2 concentrations because such changes are negligible relative to total oxygen concentration.

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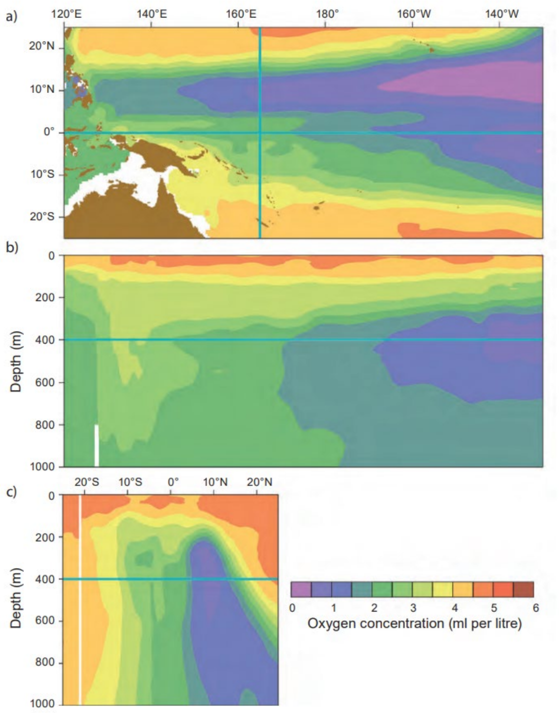

Figure 18. Dissolved oxygen. (a) Average dissolved oxygen (O_2) concentration (ml per litre) in the tropical Pacific Ocean at a depth of 400 m; and average O2 concentrations for vertical sections of the water column to depths of 1000 m at (b) the equator, and (c) 165°E. The positions of the vertical sections are indicated as blue lines in (a); horizontal blue lines in (b) and (c) represent values at a depth of 400 m as shown on (a).[100] White areas in (b) and (c) correspond to topography.[101]

Ocean acidification

The acidity of the open ocean has been relatively stable for millions of years. Due to this stability, carbonate ions (CO_3^{2-}) are so naturally abundant that the common pure minerals of calcium carbonate ($CaCO_3$) in the ocean (aragonite and calcite) are formed in surface waters and do not dissolve. This availability of carbonate ions is important to corals and other calcifying organisms that build the reefs that support coastal fisheries.[102,103] Carbonate ion availability is also important to a

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range of organisms in the food webs for tuna and related pelagic fish[104], and for many of the invertebrates that are collected for food and income by villagers throughout the tropical Pacific.[105] These organisms extract calcium (Ca^{2+}) and CO_3^{2-} from sea water to secrete the $CaCO_3$ they use to build their skeletons and shells. The stability of ocean acidity and the supply of carbonate ions are likely to be threatened by increasing levels of anthropogenic CO_2. The problem is that the increased concentration of CO_2, much of which dissolves in the ocean (Figure 19), is changing the chemistry of sea water. The consequence is that less CO_3^{2-} is expected to be available for use by calcifying organisms, reducing their growth and their chances of survival. Although the responses of species are likely to vary, depending on their physiology and the composition of their skeletal material (aragonite, calcite)[106,107], there is serious concern that ocean acidification will cause difficulties for the growth of corals and the maintenance of the essential habitats they provide for coastal fisheries. A decrease of 0.3 units in oceanic pH is expected to inhibit formation, or limit the growth, of many marine organisms.[108,109,110,111] If CO_3^{2-} declines sufficiently, aragonite (the most common form of calcium carbonate used by marine species) actually begins to dissolve.[112,113,114]

Organisms whose growth is naturally limited by the availability of CO_2, such as some phytoplankton and benthic microalgae, may benefit from rising levels of CO_2 in sea water. For example, the cryptic, bio-eroding microalgae on coral reefs grow faster under elevated CO_2 concentrations, which exacerbate reef erosion through chemical dissolution.[115] Chemical dissolution in reef carbonate sediments can also increase due to enhanced bacterial activity under higher CO_2 levels.[116,117,118]

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Average pH of ocean water is about 8.1, but it varies both seasonally and spatially by ~ 0.3 units across the oceans of the world due mainly to changes in SST and the upwelling of deep waters rich in CO_2 in the open ocean. Higher temperatures reduce the amount of CO_2 that can be dissolved in sea water, so that where the ocean is warmer, CO_2 is released and pH increases.[119] Upwelling affects pH because concentrations of CO_2 are higher in deep water due to the remineralisation of the organic matter that accumulates there. Where deep water upwells, for example, in the PEQD, the pH of surface waters is reduced. This process is also affected by seasonal and spatial changes in biological productivity.[120]

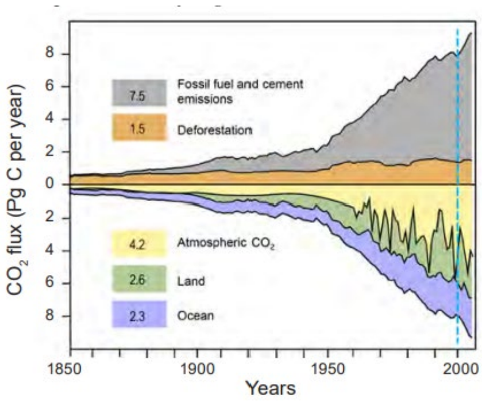

Figure 19. Sources of anthropogenic emissions of carbon dioxide (CO_2) and the 'sinks' that absorb CO_2, including the atmosphere. The flux is expressed in 10^{15}g of carbon (C) per year (Pg C per year); numbers in the colour legend represent the 2000–2007 average of the yearly flux for each component of the anthropogenic emissions (the period to the right of the dashed blue line).[121]

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