Chapter 5. Sea-Air Interactions

Contributors: Jeremy T. Mathis (Convenor), Jose Santos, Renzo Mosetti, Alberto Mavume, Craig Stevens, Regina Rodrigues, Alberto Piola, Chris Reason, Patricio A. Bernal (Co-Lead member), Lorna Inniss (Co-Lead member)

1. Introduction

From the physical point of view, the interaction between these two turbulent fluids, the ocean and the atmosphere, is a complex, highly nonlinear process, fundamental to the motions of both. The winds blowing over the surface of the ocean transfer momentum and mechanical energy to the water, generating waves and currents. The ocean in turn

he long term, the convergence and divergence of oceanic heat transport provide ces and sinks of heat for the atmosphere and partly shape the mean climate of the n. Analyzing whether these processes are changing due to anthropogenic influences the potential impact of these changes is the subject of this chapter. Following ance from the Ad Hoc Working Group of the Whole, much of the information ented here is based on or derives from the very thorough analysis conducted by the governmental Panel on Climate Change (IPCC) for its recent Fifth Assessment ort (AR5).

atmosphere and the ocean form a coupled system, exchanging at the sea-air face gases, water (and water vapour), particles, momentum and energy. These nanges affect the biology, the chemistry and the physics of the ocean and influence iogeochemical processes, weather and climate (changes affecting the water cycle addressed in Chapter 4)

n a biogeochemical point of view, gas and chemical exchanges between the oceans the atmosphere are important to life processes. Half of the Global Net Primary luction of the world is by phytoplankton and other marine plants, uptaking CO_2 and asing oxygen (

Figure 1. Global annual average sea surface temperature (SST) and Night Marine Air Temperature (NMAT) relative to a 1961–1990 climatology from state of the art data sets. Spatially interpolated products are shown by solid lines; non-interpolated products by dashed lines. From Hartmann et al. 2013, Fig. 2.18.

"It is certain that global average sea surface temperatures (SSTs) have increased since the beginning of the 20th century. (…) Intercomparisons of new SST data records obtained by different measurement methods, including satellite data, have resulted in better understanding of uncertainties and biases in the records. Although these innovations have helped highlight and quantify uncertainties and affect our understanding of the character of changes since the mid-20th century, they do not alter the conclusion that global SSTs have increased both since the 1950s and since the late 19th century." (Hartmann et al., 2013.)

2.2 Changes in sea-surface temperature (SST) as inferred from subsurface measurements.

Upper ocean temperature (hence heat content) varies over multiple time scales, including seasonal, interannual (e.g., associated with El Niño), decadal and centennial (Rhein et al., 2013). Depth-averaged (0 to 700 m) ocean-temperature trends from 1971 to 2010 are positive over most of the globe. The warming is more prominent in the Northern Hemisphere, especially in the North Atlantic. This result holds true in different analyses, using different time periods, bias corrections and data sources (e.g., with or without XBT or MBT data[1]) (Rhein et al. 2013). Zonally averaged upper-ocean temperature trends show warming at nearly all latitudes and depths (Figure 2a). However, the greater volume of the Southern Hemisphere ocean increases the contribution of its warming to the global heat content (Rhein et al., 2013). Strongest warming is found closest to the sea surface, and the near-surface trends are consistent

[1] XBT are expendable bathythermographs, probes that using electronic solid-state transducers register temperature and pressure while they free fall through the water column. MBT are their mechanical predecessors, that lowered on a wire suspended from a ship, used a metallic thermocouple as transducer.

with independently measured SST (Hartmann et al., 2013). The global average warming over this period is 0.11 [0.09 to 0.13] °C per decade in the upper 75 m, decreasing to 0.015°C per decade by 700 m (Figure 2c) (Rhein et al 2013)

The globally averaged temperature difference between the ocean surface and 200 m increased by about 0.25°C from 1971 to 2010. This change, which corresponds to a 4 per cent increase in density stratification, is widespread in all the oceans north of about 40°S. Increased stratification will potentially diminish the exchanges between the interior and the surface layers of the ocean; this will limit, for example, the input of nutrients from below into the illuminated surface layer and of oxygen from above into the deeper layers. These changes might in turn result in reduced productivity and increased anoxic waters in many regions of the world ocean (Capotondi et al., 2012)

The boundaries and names shown and the designations used on this map do not imply official endorsement or acceptance by the United Nations.

Figure 2. (a) Depth-averaged (0 to 700) m ocean-temperature trend for 1971-2010 (longitude vs. latitude, colours and grey contours in degrees Celsius per decade). (b) Zonally averaged temperature trends (latitude vs. depth, colours and grey contours in degree Celsius per decade) for 1971-2010 with zonally averaged mean temperature over-plotted (black contours in degrees Celsius). Both North (25-65°N) and South (south of 30°S), the zonally averaged warming signals extend to 700 m and are consistent with poleward displacement of the mean temperature field. Zonally averaged upper-ocean temperature trends show warming at nearly all latitudes and depths (Figure 2(b)). A relative maximum in warming appears south of 30°S (c) Globally averaged temper

2.3 Upper Ocean Heat Content (UOHC)

The ocean's large mass and high heat capacity allow it to store huge amounts of energy: more than 1000 times that found in the atmosphere for an equivalent increase in temperature. The earth is absorbing more heat than it is emitting back into space, and nearly all this excess heat is entering the ocean and being stored there.

The upper ocean (0 to 700 m) heat content increased during the 40-year period from 1971 to 2010. Published rates range from 74 TW to 137 TW (1 TW = 10^{12} watts), while

2.4 The ocean’s role in heat transport

strong impact on the northern hemisphere climate (Cunningham et al., 2013; Buchan et al., 2014).

2.5 Air-sea Heat fluxes

Heat uptake by the ocean can be substantially altered by natural oscillations in the earth's ocean and atmosphere. The effects of these large-scale climate oscillations are often felt around the world, leading to the rearrangement of wind and precipitation patterns, which in turn substantially affect regional weather, sometimes with devastating consequences.

The ENSO is the most prominent of these oscillations and is characterized by an anomalous warming and cooling of the central-eastern equatorial Pacific. The warm phase is called El Niño and the cold, La Niña. During El Niño events, a weakening of the Pacific trade winds decreases the upwelling of cold waters in the eastern equatorial Pacific and allows warm surface water that generally accumulates in the western Pacific to flow east.

As a consequence, El Niños release heat into the atmosphere, causing an increase in globally averaged air temperature. However, the "recharge oscillator theory" (Ren and Jin, 2013) indicates that a buildup of upper-ocean heat content is a necessary precondition for the development of El Niño events. La Niñas are associated with a strengthening of the trade winds, which leads to a strong upwelling of cold subsurface water in the eastern Pacific. In this case, the ocean uptake of heat from the atmosphere is enhanced, causing the global average surface temperature to decrease (Roemmich and Gilson, 2011).

The cycling of ENSO between El Niño and La Niña is irregular. In some decades El Niño has dominated and in other decades La Niña has been more frequent, also seen in phase shifts of the Interdecadal Pacific Oscillation (Meehl et al., 2013), which is related to build up and release of heat. A strengthening of the Pacific trade winds in the past two decades has led to a more frequent occurrence of La Niñas (England et al., 2014). Consequently, the heat uptake by the subsurface ocean was enhanced, leading to a slowdown of the surface warming (Kosaka and Xie, 2013). This is one of the factors affecting the global mean temperature, expected to increase by 0.21°C per decade from 1998 to 2012, but which instead warmed by just 0.04°C (the so-called recent warming hiatus, IPCC, 2013). Although there are several hypotheses on the cause of the global warming hiatus, the role of ocean circulation in this negative feedback is certain. Drijfhout et al. (2014) have shown that the North Atlantic, Southern Ocean and Tropical taking place in the Atlantic Ocean and in the Circumpolar Current region. Coinciding in time, changes in OHC could help to explain the observed slowdown in global warming. It is anticipated that the mechanisms involved may at some point reverse, releasing large amounts of heat to the atmosphere and accelerating global warming (e.g., Levermann, et al., 2012).

Many other naturally occurring ocean-atmosphere oscillations in the Pacific, Atlantic, and Indian Oceans have also been recognized and named. The ENSO, as a global phenomenon, has an expression in the Atlantic basin called the Atlantic Niño. In the last six decades, this mode has weakened, leading to a warming of the equatorial eastern

of America suffer droughts. La Niña events usually cause the opposite patterns. However, in the last several decades, ENSO events have changed their spatial and temporal characteristics (Yeh et al., 2009; McPhaden, 2012).

During recent decades, the warm waters of El Niño events have been displaced to the central Pacific instead of to the eastern Pacific. It is not clear yet whether these changes are linked to anthropogenic climate change or natural variability (Yeh et al., 2011). In any case, the effects on climate of an ENSO event centred in the central Pacific (a central Pacific ENSO) are in sharp contrast to that associated with one centred in the eastern Pacific.

For instance, northeastern and southeastern Australia experience a reduction in rainfall during the eastern Pacific El Niños and there is a decrease in rainfall over northwestern and northern Australia during central Pacific events (Taschetto and England, 2009; Taschetto et al., 2009). The Indian monsoon fails during eastern Pacific El Niños, but is enhanced during central Pacific El Niños (Kumar et al., 2006). Over the semi-arid region of northeast Brazil, eastern Pacific El Niños/La Niñas cause dry/wet conditions; central Pacific El Niños have the opposite effect, with the worst drought in the last 50 years associated with the strong 2011/12 La Niña and not with El Niños as in the past (Rodrigues et al., 2011; Rodrigues and McPhaden, 2014). This drought caused the displacement of 10 million people and economic losses on the order of 3 billion United States dollars in relation to agriculture and cattle raising alone. In contrast to drought in Brazil, the 2011/12 La Niña caused floods across southeastern Australia.

In other ocean basins, changes in oceanic oscillations and temperatures have also had an impact on climate. For instance, in the Indian Ocean, a positive phase of the Indian Dipole Mode (warm/cold temperatures in the western/eastern equatorial Indian Ocean) leads to flooding in east 0(o)-bt tia p /P <</;dtatotii12(e)-ljdonein3e Io (e)-lj2(s)6(e)-1(as)2(t)10(M 1999; Ashok et al., 2001; Gadgil et al., 2004; Yamagata et al., 2004; Behera et al., 2005; Ummenhofer et al., 2009; Cai et al., 2011). The counterpart of ENSO in the Atlantic (Atlanti-4( )10(N)-3(iñ)6(o)-2())4(h)-4(as)2(w)6.1(e)-1(ak)16(e)-1(n)-4(e)-1(d)6(d)6(u)-4(rin)-4(g) the eastern equatorial Atlantic. As a consequence, rainfall4( )10(ha)4(s)16(be)13(e3(e)ua)4((r)

The IPCC AR5 concluded that “it is unlikely that annual numbers of tropical storms, hurricanes and major hurricanes counts have increased over the past 100 years in the North Atlantic basin. Evidence, however, is for a virtually certain increase in the human time scales, seawater's salinity can only be altered—over days or centuries—by the addition or removal of fresh water.

The water cycle is expected to intensify in a warmer climate. Observations since the 1970s show increases in surface and lower atmospheric water vapour (Figure 4a), at a rate consistent with observed warming. Moreover, evaporation and precipitation are projected to intensify in a warmer climate. Recorded changes in ocean salinity in the last 50 years support that projection (Rhein et al. 2013; FAQ. 3.2).

The atmosphere connects the ocean's regions of net fresh water loss to those of fresh water gain by moving evaporated water vapour from one place to another. The distribution of salinity at the ocean surface largely reflects the spatial pattern of evaporation minus precipitation (Figure 4b), runoff from land, and sea ice processes. There is some shifting of the patterns relative to each other, because of the ocean's currents. Ocean salinity acts as a sensitive and effective rain gauge over the ocean. It naturally reflects and smoothes out the difference between water gained by the ocean from precipitation, and water lost by the ocean through evaporation, both of which are very patchy and episodic (Rhein et al. 2013; FAQ 3.2). Data from the past 50 years show widespread salinity changes in the upper ocean, which are indicative of systematic changes in precipitation and runoff minus evaporation.

(Figure 4b). Subtropical waters are highly saline, because evaporation exceeds rainfall, whereas seawater at high latitudes and in the tropics—where more rain falls than evaporates—is less so. The Atlantic, the saltiest ocean basin, loses more freshwater through evaporation than it gains from precipitation, while the Pacific is nearly neutral, i.e., precipitation gain nearly balances evaporation loss, and the Southern Ocean is dominated by precipitation. (Figure 4b; Rhein et al. 2013; FAQ. 3.2). Changes in surface salinity and in the upper ocean have reinforced the mean salinity pattern (4c). The evaporation-dominated subtropical regions have become saltier, while the precipitation-dominated subpolar and tropical regions have become fresher. When changes over the top 500 m are considered, the evaporation-dominated Atlantic has become saltier, while the nearly neutral Pacific and precipitation-dominated Southern Ocean have become fresher (Figure 4d; Rhein et al. 2013; FAQ. 3.2)

Observed surface salinity changes also suggest a change in the global water cycle has occurred (Chapter 4). The long-term trends show a strong positive correlation between the mean climate of the surface salinity and the temporal changes in surface salinity from 1950 to 2000. This correlation shows an enhancement of the climatological salinity pattern: fresh areas have become fresher and salty areas saltier.

Ocean salinity is also affected by water runoff from the continents, and by the melting and freezing of sea ice or floating glacial ice. Fresh water added by melting ice on land will change global-averaged salinity, but changes to date are too small to observe (

The boundaries and names shown and the designations used on this map do not imply official endorsement or acceptance by the United Nations.

Figure 4. Changes in sea surface salinity are related to the atmospheric patterns of evaporation minus precipitation (E – P) and trends in total precipitable water: (a) Linear trend (1988–2010) in total precipitable water (water vapour integrated from the Earth's surface up through the entire atmosphere) (kg m^{-2} per decade) from satellite observations (Special Sensor Microwave Imager) (after Wentz et al., 2007) (blues: wetter; yellows: drier). (b) The 1979–2005 climatological mean net E – P (cm yr^{-1}) from meteorological reanalysis (National Centers for Environmental Prediction/National Center for Atmospheric Research; Kalnay et al., 1996) (reds: net evaporation; blues: net precipitation). (c) Trend (1950–2000) in surface salinity (PSS78 per 50 years) (after Durack and Wijffels, 2010) (blues freshening; yellows-reds saltier). (d) The climatological-mean surface salinity (PSS78) (blues: <35; yellows–reds: >35). From Rhein et al. 2013; FAQ. 3.2. Fig 1.

In conclusion, according to the last IPCC AR5, "it is very likely that regional trends have enhanced the mean geographical contrasts in sea surface salinity since the 1950s: saline

Figure 5. CO_2 emissions from different sources from 1958 to 2013 (Le Quéré et al. 2014)

Coal is an important and, recently, growing proportion of CO_2 emissions from fossil fuel combustion. From 2012 to 2013, CO_2 emissions from coal increased 3.0 per cent, compared to the increase rate of 1.4 per cent for oil and gas (Le Quéré et al. 2014). Coal accounted for about 60 per cent of the CO_2 emission growth in the same period. This is largely because many large economies of the world have recently resorted to using coal as an energy source for a wide variety of industrial processes, instead of a [(r)4(C)8(O)]TJ 0 Tc 59 2

The boundaries and names shown and the designations used on this map do not imply official endorsement or acceptance by the United Nations.

Figure 6. Anthropogenic CO_2 distributions along representative meridional sections in the Atlantic, Pacific, and Indian oceans for the mid-1990s (Sabine et al. 2004).

Because the ocean mixes slowly, about half of the anthropogenic CO_2 (C_{ant}) stored in the ocean is found in the upper 10 per cent of the ocean (Figure 6). On average, the penetration depth is about 1000 meters and about 50 per cent of the anthropogenic CO_2 in the ocean is shallower than 400 meters.

Globally, the ocean shows large spatial variations in terms of its role as a sink of atmospheric CO_2 (Takahashi et al. 2009). Over the past 200 years the oceans have absorbed 525 billion tons of CO_2 from the atmosphere, or nearly half of the fossil fuel emissions over the period (Feely et al. 2009). The oceanic sink of atmospheric CO_2 has increased from 4.0 ± 1.8 $GtCO_2$ (1 $GtCO_2$ = 10^9 tons of carbon dioxide) per year in the 1960s to 9.5 ± 1.8 $GtCO_2$ per year during 2004-2013. During the same period, the estimated annual atmospheric CO_2 captured by the ocean was 2.6 ± 0.5 Gt of CO_2 compared with around 19 Gt of CO_2 during the sixties (Le Queré et al. 2014). However, due to the decreased buffering capacity, caused by this CO_2 uptake, the proportion of anthropogenic carbon dioxide that goes into the ocean has been decreasing.

Estimates of the global inventory of anthropogenic carbon (C_{ant}) (including marginal seas) have a mean value of 118 PgC and a range of 93 to 137 PgC in 1994 and a mean of 160 PgC and range of 134 touC inu2me of o89 0 Td ( )Tj -0.005 Tc 0.if 1632(nn632(n 3d10()4(e)11(c

The storage rate of anthropogenic CO_2 is assessed by calculating the change in C_{ant} concentrations between two time periods. Regional observations of the storage rate are in general agreement with that expected from the increase in atmospheric CO_2 concentrations and with the tracer-based estimates. However, there are significant spatial and temporal variations in the degree to which the inventory of C_{ant} tracks changes in the atmosphere (Figure 7, Rhein et al 2013)

Although the average oceanic pH can vary on interglacial time scales, the changes are usually on the order of ~0.002 units per 100 years; however, the current observed rate of change is ~0.1 units per 100 years, or roughly 50 times faster. Regional factors, such as coastal upwelling, changes in riverine and glacial discharge rates, and sea-ice loss have created “OA hotspots” where changes are occurring at even faster rates. Although OA is a global phenomenon that will likely have far-reaching implications for many marine organisms, some areas will be affected sooner and to a greater degree.

Recent observations show that one such area in particular is the cold, highly productive region of the subarctic Pacific and western Arctic Ocean, where unique biogeochemical processes create an environment that is both sensitive and particularly susceptible to accelerated reductions in pH and carbonate mineral concentrations. The OA phenomenon can cause waters to become undersaturated in carbonate minerals and thereby affect extensive and diverse populations of marine calcifiers.

4.4 The CO_2 problem

Based on the most recent data of 2004 to 2013, 35.7 $GtCO_2$ (1 $GtCO_2$ = 10^9 tons of carbon dioxide) of anthropogenic CO_2 are released into the atmosphere every year (Le Quéré et al. 2014). Of this, approximately 32.4 $GtCO_2$ come directly from the burning of fossil fuels and other industrial processes that emit CO_2. The remaining 3.3 $GtCO_2$ are due to changes in land-use practices, such as deforestation and urbanization. Of this 35.7 $GtCO_2$ of anthropogenically produced CO_2 emitted annually, approximately 10.6 $GtCO_2$ (or 29 per cent) are incorporated into terrestrial plant matter. Another 15.8 $GtCO_2$ (or 46 percent) are retained in the atmosphere, which has led to some planetary warming. The remaining 9.5 $GtCO_2$ (or 26 per cent) are absorbed by the world’s oceans (Le Quéré et al. 2014).

As the hydrogen ions produced by the increased CO_2 dissolution take carbonate ions out of seawater, the rate of calcification of shell-building organisms is affected; they are confronted with additional physiological challenges to maintain their shells. Although alteration of the carbonate equilibrium system in the ocean reducing carbonate ion concentration, and saturation states of calcium carbonate minerals will play a role imposing an additional energy cost to calcifier organisms, such as corals and shellbearing plankton, this is by no means the sole impact of OA.

4.5 What are the impacts of a more acidic ocean?

Throughout the last 25 million years, the average pH of the ocean has remained fairly constant between 8.0 and 8.2. However, in the last three decades, a fast drop has begun to occur, and if CO_2 emissions are left unchecked, the average pH could fall below 7.8 by the end of this century (Rhein, et al. 2013).

This is well outside the range of pH change of any other time in recent geological history. Calcifying organisms in particular, such as corals, crabs, clams, oysters and the tiny free-swimming pteropods that form calcium carbonate shells, could be particularly vulnerable, especially during the larval stage. Many of the processes that cause OA have long been recognized, but the ecological implications of the associated chemical

References

Abraham, J.P., Baringer, M., Bindoff, N.L., Boyer, T., Cheng, L.J., Church, J.A., Conroy, J.L., Domingues, C.M., Fasullo, J.T., Gilson, J., Goni, G., Good, S.A., Gorman, J.M., Gouretski, V., Ishii, M., Johnson, G.C., Kizu, S., Lyman, J.M., Macdonald, A.M., Minkowycz, W.J., Moffitt, S. E., Palmer, M.D., Piola, A.R., Reseghetti, F., Schuckmann, K., Trenberth, K.E., Velicogna, I., & Willis, J.K. (2013). A review of global ocean temperature observations: Implications for ocean heat content estimates and climate change, Review of Geophysics, 51, 450–483, doi:10.1002/rog.20022.

Ashok K., Guan, Z., Yamagata, T. (2001), Impact of the Indian Ocean dipole on the relationship between the Indian monsoon rainfall and ENSO, Geophysical Research Letters, 28, 4499–4502.

Behera S. K., Luo, J.-J., Masson, S., Delecluse, P., Gualdi, S., Navarra, A., Yamagata, T. (2005), Paramount impact of the Indian Ocean dipole on the east African short rains: a CGCM study. Journal of Climate, 18, 4514–4530.

Bresnan, E., Davidson, K., Edwards, M., Fernand, L., Gowen, R., Hall, A., Kennington, K., McKinney, A., Milligan, S., Raine, R., Silke, J. (2013) Impacts of climate change on harmful algal blooms, Cunningham, S. A., Kanzow, T., Rayner, D., Baringer, M.O., Johns, W.E., Marotzke, J., Longworth, H.R., Grant, E.M., Hirschi, J.J.-M., Beal, L.M., Meinen, C.S., Bryden, H.L. (2007), Temporal variability of the Atlantic meridional overturning circulation at 26.5 °N, Science, 317, 935–938, doi: 10.1126/science.1141304.

Cunningham, S. A., Roberts, C.D., Frajka-Williams, E., Johns, W.E., Hobbs, W., Palmer, M.D., Rayner, D., Smeed, D.A., and McCarthy, G. (2013), Atlantic Meridional Overturning Circulation slowdown cooled the subtropical ocean, Geophysical Research Letters, 40, 6202–6207, doi:10.1002/2013GL058464.

Doney S.C., Ruckelshaus M., Duffy J.E., Barry J.P., Chan F., English C.A., Galindo H.M., Grebmeier J.M., Hollowed, A.B., Knowlton, N., Polovina, J., Rabalais, N.N., Sydeman, W.J., Talley, L.D. (2012). Climate change impacts on marine ecosystems. Annual Review of Marine Science 4:11-37.

Doney, S.C., Mahowald, N., Lima, I., Feely, R.A., Mackenzie, F.T., Lamarque, J.-F. and Rasch, P.J. (2007). Impact of anthropogenic atmospheric nitrogen and sulfur
I-P., 0 Tc 0 Tw 1.78 00Td ( )Tj 0.005 Tc -0.001 Tw50.23 0 Td [(R)4(WN)]h(ti)4(,)]TJ 5TJ -

Field, C.B., Behrenfeld, M.J., Randerson, J.T. and Falkowski, P. (1998). Primary production of the biosphere: integrating terrestrial and oceanic components, Science 281, 237 Falko57

Levermann, A., Bamber, J.L., Drijfhout, S., Ganopolski, A., Haeberli, W., Harris, N.R.P., Huss, M., Krüger, K., Lenton, T.M., Lindsay, R.W., Notz, D., Wadhams, P. and Weber, S. (2012). Potential climatic transitions with profound impact on Europe: Review of the current state of six 'tipping elements of the climate system'. Climatic Change, 110, 845–878, DOI 10.1007/s10584-011-0126-5.

Levitus, S., Antonov, J.I., Boyer, T.P., Locarnini, R.A., Garcia, H.E., and Mishonov, A.V. (2009). Global ocean heat content 1955–2008 in light of recently revealed instrumentation problems, Geophysical Research Letters, 36, L07608, doi:10.1029/2008GL037155.

Lima, F.P., and Wethey, D.S. (2012). Three decades of high-resolution coastal sea surface temperatures reveal more than warming, Nature communications, 3.

Marengo, J. A., Tomasella, J., Alves, L.M., Soares, W.R., and Rodriguez, D.A. (2011), The drought of 2010 in the context of historical droughts in the Amazon region, Geophysical Research Letters, 38, L12703, doi:10.1029/2011GL047436.

oM.-7(J-2

Suthers, I.M., Young, J.W., Baird, M.E., Roughan, M., Everett, J.D., Brassington, G.B., Byrne, M., Condie, S.A., Hartog, J.R., Hassler, C.S., Hobday, A.J., Holbrook, N.J.,